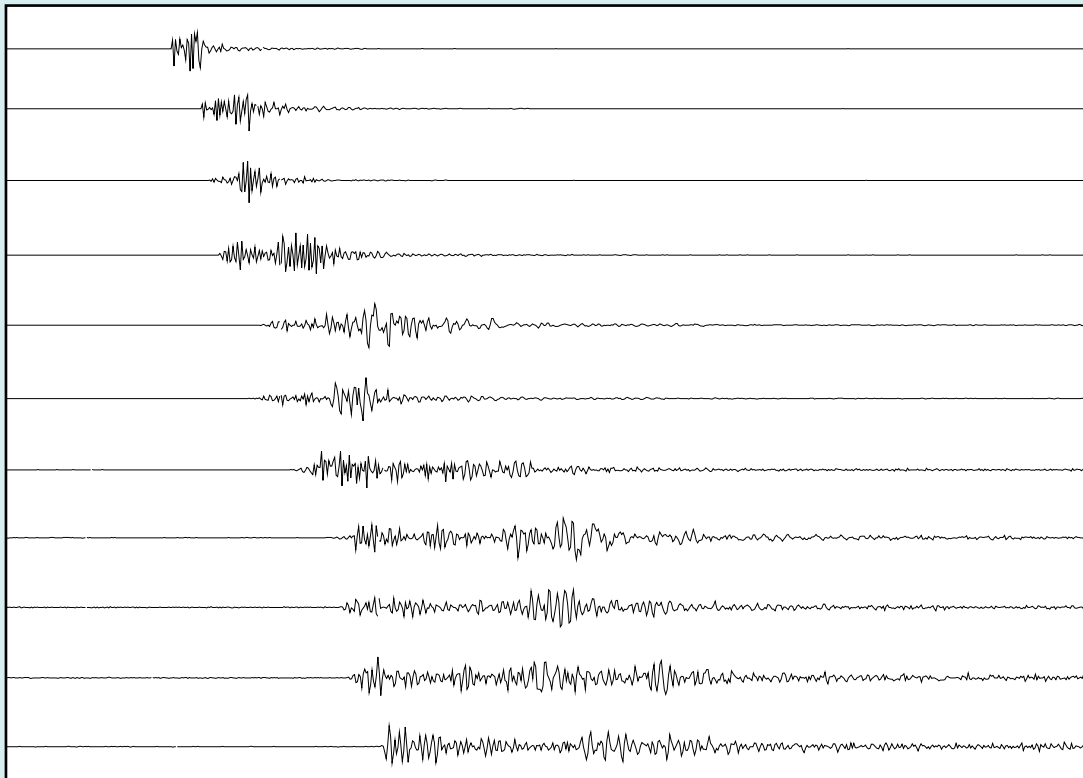


Hydrotectonics and fault criticality analysis in karstic regions from microseismic and hydrogeological observations



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*"If the sky that we look upon
Should tumble and fall
Or the mountain should crumble to the sea
I won't cry [...]
No, I won't shed a tear
Just as long as you stand,
Stand by me"*

Ben. E. King 1961

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Abstract

This study aims at assessing fault criticality in karstic regions through the analysis of hydrogeological and microseismic data, at enhancing the understanding of interactions between seismic activity and groundwater circulation and using this insight to estimate the present-day stress state of the shallow crust. These objectives are prompted by the present concerns regarding induced seismicity and its associated risks in projects involving for example deep geothermal energy exploitation, CO₂ sequestration or radioactive waste disposal. In the context of induced seismicity, fluids play a major role by influencing the effective stresses and thus fault stability. While human-induced seismicity is often observed in cases of waste water injection or reservoir stimulation, naturally triggered seismicity following seasonal recharge can also be observed, especially in karstic areas where the channeling effect of karst conduits has a major influence on pore pressure. Thus, combining natural seismicity and hydrogeological data can inform on the interaction between fluid flow and fault rupture. To fulfill these goals, the initial phase involves the identification and the quantification of what mechanisms may trigger seismicity following an important recharge period (heavy precipitation or snow melt) in karstic regions. The identified mechanisms are 1) an increasing vertical load due to the additional water within the karst conduits, 2) poro-elastic deformation resulting from the additional load, 3) a massive pore pressure increase resulting from a direct hydrogeological connection between the infiltrated water and focal depth and 4) pressure diffusion process. These mechanisms, whether they act alone or together, influence the pore pressure, however the timing of their effectiveness varies. The response of the pore pressure to an increasing hydraulic head is instantaneous for mechanism 1-3 but delayed for 4. Using data from 3 test sites and 1D analytical solutions to calculate pore pressure response to an increasing hydraulic head allows to determine, which process is more likely to trigger seismicity following heavy rainfalls. In parallel to this initial phase, a seismic and hydrogeological monitoring network is deployed in the Neuchâtel Jura to create a seismic catalog and collect continuous flow rates from major karstic springs, which is used as proxy for groundwater level variations. These two data sets are then combined and a statistically significant link is found between high groundwater conditions and seismic activity. A detailed analysis of the seismic activity of the analyzed fault zones and their associated rain-triggered events, coupled with the afore-mentioned 1D analytical solutions is then performed. For each fault zone a triggering pore pressure increase is defined and used to determine the sensitivity of the faults to hydraulic head variations. Although the majority of the investigated faults show some activity following important precipitations, events do not occur on a regular basis. The identification of rain-triggered events, the detection threshold of seismic events as well as hydromechanical properties of the faults may explain this lack of consistency. Integrating

this knowledge, together with the overall seismic activity of the region and the geological and hydrogeological context, a conceptual hydromechanical model for the study area is presented. Finally, the state of stress of the shallow crust is estimated using two distinct methods and a range of values and orientations is presented. As a conclusion, this study mainly stresses the importance of considering the hydrogeological context when discussing rain-triggered seismicity, as the channeling effect of karst conduits drastically increases the pore pressure at depth, and consequently has a major influence on the magnitude of the triggering pore pressure. This study also contributes to a higher understanding of the seismic activity near Lake Neuchâtel and how it is influenced by the geological and hydrogeological context.

Keywords

Fault criticality, Rain-triggered seismicity, Karst, Pore-pressure diffusion

Résumé

Cette étude vise à évaluer la criticalité des failles dans les régions karstiques en combinant l'analyse de données hydrogéologiques et microsismiques, à améliorer la compréhension des interactions entre l'activité sismique et la circulation des eaux souterraines et à utiliser ces connaissances pour estimer l'état de contrainte actuel de la croûte supérieure. Ces objectifs sont motivés par les préoccupations actuelles concernant la sismicité induite et les risques qui y sont associés dans des projets tels que l'exploitation de l'énergie géothermique profonde, la séquestration de CO₂ ou encore l'enfouissement des déchets radioactifs. Dans le contexte de la sismicité induite, l'eau joue un rôle majeur en influençant les contraintes effectives, et donc la stabilité des failles. Si la sismicité induite par l'Homme est souvent observée dans les cas d'injection d'eaux usées ou de stimulation de réservoirs, la sismicité induite naturellement peut également être observée à la suite d'intense période de recharge, en particulier en zone karstique où l'effet de canalisation des conduits karstiques a une influence majeure sur la pression interstitielle. Ainsi, en combinant des données hydrogéologiques aux données de sismicité naturelle, il est possible d'obtenir des informations sur l'interaction de la circulation des fluides et la stabilité des failles. Pour atteindre ces objectifs, la phase initiale du projet consiste à identifier et à quantifier les mécanismes susceptibles de déclencher un séisme à la suite d'une recharge importante (événement pluvieux ou fonte des neiges) dans les régions karstiques. Les mécanismes identifiés sont 1) une augmentation de la charge verticale due à l'eau supplémentaire dans les conduits karstiques, 2) une augmentation de la pression interstitielle suite à une déformation poro-élastique résultant de la charge supplémentaire, 3) une augmentation massive de la pression interstitielle résultant de la connexion hydrogéologique directe entre l'eau infiltrée et la profondeur focale et 4) un processus de diffusion de pression. Ces mécanismes, qu'ils agissent seuls ou ensemble, influencent la pression interstitielle, mais le moment auquel ils agissent varie. La réponse de la pression suite à une augmentation de la charge hydraulique est instantanée pour les mécanismes 1 à 3, mais retardée pour le mécanisme 4. L'utilisation de données provenant de 3 études de cas ainsi que des solutions analytiques 1D permettant de calculer la variation de la pression interstitielle suite à une augmentation massive de la charge hydraulique permet d'identifier quel processus est le plus susceptible de générer un séisme à la suite de fortes précipitations. En parallèle à cette première phase du projet, un réseau de surveillance sismique et hydrogéologique est déployé dans le Jura Neuchâtelois afin de créer un catalogue sismique et de collecter les débits des principales sources karstiques de la région, utilisés comme indicateurs des variations du niveau des eaux souterraines. Ces deux ensembles de données sont ensuite combinés et un lien statistiquement significatif est trouvé entre des périodes de hautes eaux souterraines et l'activité sismique. Une analyse détaillée de l'activité sismique des différentes zones de faille étudiées ainsi

que des événements induits par la pluie qui leur sont associés est réalisée. En utilisant les solutions analytiques mentionnées ci-dessus, les variations de pressions interstitielles sont calculées pour chaque zone de faille au moment de la rupture et une valeur seuil au-delà de laquelle un séisme est envisageable, est définie pour chaque faille. Bien que la majorité des failles étudiées montre une certaine activité à la suite d'intenses précipitations, les tremblements de terre ne se produisent pas de manière régulière. L'identification des événements induits par la pluie, le seuil de détection des tremblements de terre ou encore les propriétés hydromécaniques des failles peuvent expliquer ce manque de cohérence. En intégrant ces connaissances au contexte géologique et hydrogéologique de la région, un modèle conceptuel hydromécanique est établi. Dans la dernière partie de ce travail, l'état des contraintes actuelles de la croûte superficielle est estimé à l'aide de deux méthodes distinctes et une gamme de valeurs et d'orientations, dont les résultats sont en accord avec des études antérieures, sont présentées. En conclusion, ce travail souligne l'importance de prendre en compte le contexte hydrogéologique d'une région lors de discussion sur la sismicité induite par la pluie, car l'effet de la canalisation des conduits karstiques augmente considérablement la pression interstitielle en profondeur, et a par conséquent une influence majeure sur la valeur seuil identifiée. Cette étude contribue également à une meilleure compréhension de l'activité sismique aux abords du lac de Neuchâtel et de la manière dont le contexte géologique et hydrogéologique l'influence.

Mots-clés

Criticalité des failles, Séismes induits par la pluie, Karst, Diffusion de pression

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Chapter 1

Introduction

1.1 Motivations

Better understanding fault criticality, the proximity of a fault to shear failure, is of primary interest when planning underground projects. Stress perturbations caused by human activities such as injection or extraction of fluids, in the surroundings of a critically stressed fault, can affect the fault's stability. Cases of human-induced seismicity have been reported in several places over the last decades following, for example, water injection in deep wells (Parotidis et al., 2004; Ellsworth, 2013), massive groundwater extraction (González et al., 2012) or reservoir impoundment (Gupta, 2002; Kangi and Heidari, 2008). Figure 1.1 illustrates how the underground infrastructures interact with the host rock and deformation structures. Concerns about the seismic risk resulting from human activities and the fear of a devastating earthquake is growing in society, especially for people living close to infrastructures susceptible to induce seismicity. Past experiences, such as the Basel geothermal project (Häring et al., 2008), which was abandoned due to the unacceptable seismic risk, highlight the importance of better mitigating this risk in order to avoid important financial losses (Kraft et al., 2009). Investigating the fault's stability and its sensitivity to external stress changes improves our knowledge about local stress conditions and provides valuable information for seismic risk analysis.

1.2 Context

The interaction between water and tectonic forces, regrouped under the term of hydrotectonics, plays a major role in triggered or induced seismicity. In addition to the processes mentioned above, water injection, extraction and reservoir impoundment, cases of naturally-induced seismicity have been observed (e.g. Saar and Manga (2003); Manga et al. (2016)) and include case studies reporting seasonal patterns of seismicity and their relation to groundwater recharge or seasonal snow load (Wolf et al., 1997; Bettinelli et al., 2008; Braunmiller et al., 2014; Birhanu et al., 2018). Naturally-induced seismicity following rainfalls (hereafter referred to as rain-triggered seismicity) is observed in a variety of hydrogeologic and tectonic settings and the time lag between the rainfall and seismicity varies from hours to months (Costain and Bollinger, 2010). For example, Muço (1995) observed that seismicity in Albania was well correlated with the wet season with a time lag of 1-2 months between the maximum intensity of precipitation

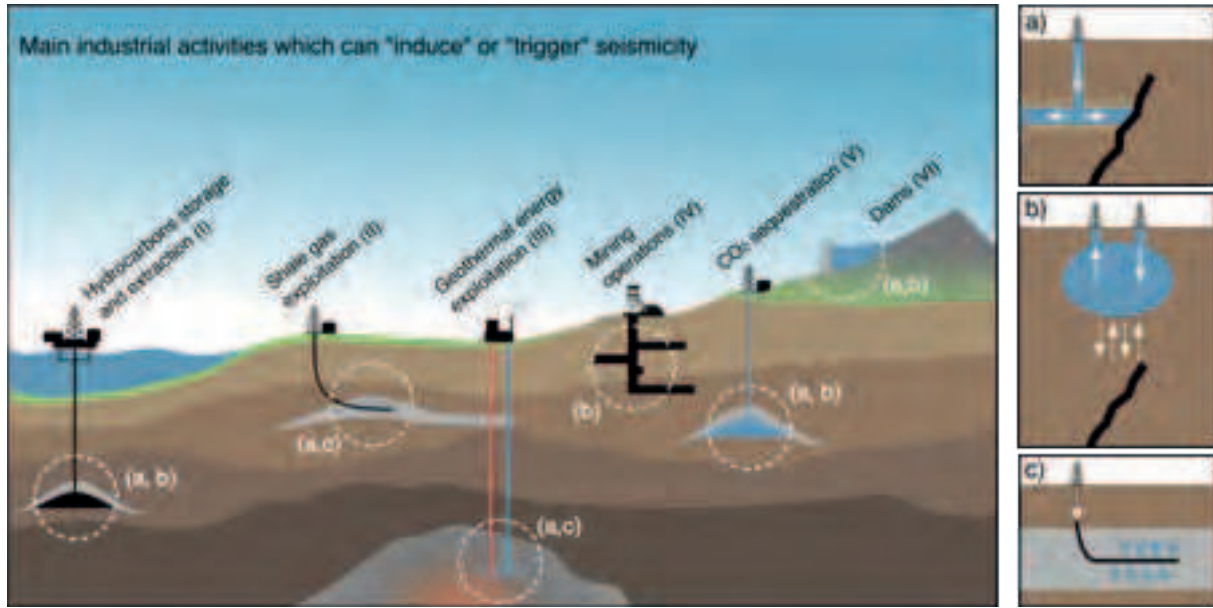


Figure 1.1: Major industrial activities that can induce seismicity. a) Earthquake triggered by fluid injection, b) Earthquake triggered by a mass or a volume change and c) Seismicity generated through hydraulic fracturing. Figure taken from www.destress-h2020.eu/en/Best-Practices/Induced-Seismicity/, August 2023, modified from [Grigoli et al. \(2017\)](#)

and the earthquake occurrence. [Bollinger et al. \(2007\)](#) observed a strong seasonal modulation of seismicity, with nearly twice as much events during the winter period in the Himalaya. This corresponds to a time lag of 6 months with the monsoon regime in summer, however, their preferred hypothesis is that the important recharge in groundwater during the summer inhibits thrusting through an increasing vertical load due to the infiltrated water. [Saar and Manga \(2003\)](#), found a time-lag of 151 days between groundwater recharge and seismicity at Mount Hood, Oregon, and inferred from this observation, together with a mean depth for seismicity, hydraulic properties of the substratum, such as hydraulic diffusivity, hydraulic conductivity and permeability. Among cases of rain-triggered seismicity, some do not show a seasonal pattern but a punctual increase in the seismic rate following important precipitations ([Kraft et al. 2006a](#); [Husen et al. 2007](#); [Rigo et al. 2008](#)). These case studies show a short time interval between the maximum rain intensity and the beginning of seismicity (hours to days) and happen to have a similar geological context: karstified limestones ([Kraft et al. 2006a](#); [Husen et al. 2007](#); [Rigo et al. 2008](#)). The primary mechanism invoked for explaining seismicity following heavy rainfalls is pore fluid pressure diffusion, similarly to what is described in reservoir induced seismicity ([Talwani, 1997](#); [Talwani and Acree, 1984](#)), where the onset of precipitation acts like a pressure front that migrates from the surface towards focal depth. Although this is a reasonable hypothesis considering a homogeneous fractured crust, it is too simplistic for a known karstic area. Moreover, seismicity strongly correlated with rainfalls is often used to determine physical properties of the crust such as hydraulic diffusivity or the state of stress of the crust ([Roth et al. 1992](#); [Saar and Manga, 2003](#); [Kraft et al. 2006a](#); [Husen et al. 2007](#); [Hainzl et al. 2006](#)). Ignoring the intrinsic properties of groundwater circulation in karstic areas can lead to erroneous parameter estimation.

Karst and groundwater circulation

In their book, [Ford and Williams \(2013\)](#) define karst as a special landscape developing where regional geology mainly consists in soluble rocks, such as limestones and evaporites. Karst development does not solely depend on the rock's composition: a heavily fractured rock, where water flow is concentrated along pre-existing fractures and/or bedding planes will easily develop karstic channels. This concentrated water flow results in a progressive erosion and develops the typical morphologies of karst such as poljé, dolines, caves and underground channels. The general morphology of a karstic network can be distinguished into 4 zones according to their respective flow conditions (epikarst, vadose, epiphreatic and phreatic) ([Palmer, 1991](#); [Audra and Palmer, 2013](#); [Jouves et al., 2017](#)). Following the description made by [Jouves et al. \(2017\)](#), the epikarst is the superficial layer of the karstic network and collects the water before draining it towards the vadose zone. The vadose zone is dominated by vertical shafts, where water flow is driven by gravity towards the epiphreatic zone, where conduits can be vertical and horizontal. The epiphreatic zone corresponds to the zone of fluctuating groundwater levels between low waters and high waters. At the bottom of the epiphreatic zone, i.e. at the limit between the epiphreatic and the phreatic zone, the conduits are mostly horizontal. Flooded conduits can also be found below the base level, in the phreatic zone, either due to variation of base level over time or due to hypogene recharge due to ascending flows. A karstic system is said to be mature when the channels are well developed in a low permeable matrix. This difference in hydraulic parameters between the channels and the matrix leads to the so-called duality of karst mentioned in the work of [Király \(1994\)](#). [Király \(1998\)](#) evokes the duality of karst as being a direct consequence of its structure (high permeable conduits and low permeable matrix), having an influence on:

- (i) Infiltration: concentrated or diffuse
- (ii) Groundwater flow: high velocities or low velocities
- (iii) Discharge: concentrated or diffuse seepage

This duality satisfactorily explains the shape of a karstic spring hydrograph following a recharge event ([Király, 1994, 1998](#)), where the hydrograph represents the combined response of both, the karst conduits and the low permeable matrix, with a rapid response from the conduits and a delayed response from the matrix. Analysis of karstic hydrograph can inform about the structures of the karstic system. From his numerical modelling, [Király \(1998\)](#) estimated that more than 40 % of the infiltration has to be concentrated to reproduce the typical karstic hydrograph, implying a high drainage capacity of the epikarst. Such important concentrated infiltration and high flow velocities exceeds the evacuation capacity of the system and leads to a rapid increase in the hydraulic head in the conduits (for example, a rate of increase greater than 2 m/h in the Cetina catchment in the Dinaric karst, Croatia ([Bonacci, 1987](#)) and up to 16 m/h in the Serrière catchment in the Swiss Jura ([Huguenin, 2020](#))). This process results in an inversion of the flow regime, during which water flows from the conduits to the matrix, recharging the matrix. During recession, water flows from the matrix to the conduits, maintaining the base flow level. Several studies on karst aquifers show that between 60 and 80% of the total quantity

of water flows through the conduits (Atkinson, 1977; Bonacci, 1987) although their volume can be 30 times smaller than the volume of the unkarstified matrix for a given karst massif (Bonacci, 1987).

Fault rupture and earthquakes

A fault ruptures when the fault's strength, i.e. the ability of the fault to resist to deformation of the surrounding rock, is overcome. Commonly, the Coulomb rupture criterion is used to determine what stress conditions may initiate a rupture on a fault (Jaeger et al., 2009):

$$\tau = c + \mu \sigma'_n \quad (1.1)$$

where τ is the shear stress on the fault plane, c the cohesion, μ the coefficient of friction and the effective normal stress $\sigma'_n = \sigma_n - p$, with σ_n the normal stress and p the pore pressure in the fault zone. When the shear stress τ is greater than the shear strength ($c + \mu \sigma'_n$), the fault ruptures. The strength of the fault is characterized by the cohesion c , which can be seen as the intrinsic ability of the material to be cohesive, i.e., to hold together, the friction coefficient μ , which relates to the internal friction angle ϕ of a material by the following relation $\mu = \tan(\phi)$ and the effective normal stress to the fault σ'_n , i.e., the stress that acts perpendicularly to the fault plane. The effective normal stress σ'_n will vary with pore pressure ($\sigma'_n = \sigma_n - p$), and decrease if the pore pressure increases, reducing the shear strength (Hubbert and Rubey, 1959). While Eq. (1.1) informs whether the fault slips or not, it does not indicate whether the slip causes an earthquake or not. When a fault slips without generating an earthquake, it is called an aseismic deformation or fault creep. Such deformation can be observed at a large scale, such as the Maacama and Calaveras Fault in California (Evans et al., 1981; Swiatlowski, 2020), but also at a decametric scale as proved by field experiments (Guglielmi et al., 2015; Cornet, 2016). The difference between a seismic (earthquake) or an aseismic (no earthquake) slip lies in friction coefficient stability. While the internal friction coefficient of an intact rock is an intrinsic property of the material, the friction coefficient during fault reactivation varies according to the roughness (surface geometry) of the fault plane and its composition (fault gouge) (Lockner and Beeler, 2002). Moreover, when a slip is initiated, the friction coefficient will change from the static friction coefficient μ_s to the dynamic friction coefficient μ_d . The change between μ_s and μ_d is achieved over a slip distance D_c and depends on the sliding velocity (Scholz, 1998; Weiss et al., 2016). μ_d can either decrease or increase with the slip, called respectively slip-weakening and slip-hardening processes (e.g. Ohnaka et al., 2013). Coupled with the sliding velocity, these effects are found within the rate-and-state friction law (Dieterich, 1979), that is widely used for the modelling of seismic cycles and gives insight on how the changes in friction may affect the stick and slip model of earthquake generation (Brace and Byerlee, 1966). It is commonly assumed that earthquake are generated through a frictional instability, i.e., variation in the frictional resistance along the surfaces, as expressed in the stick-slip model for earthquake generation, presented by Brace and Byerlee (1966): It was observed that when two surfaces of rock slide over one another, the motion is jerky, i.e, it will slip over a certain distance and then stick for a while before slipping again. Tests were performed on rock samples placed

under a confining pressure, which were loaded on both ends under compression. Each time the sample slipped along the fracture, a stress drop was observed, directly followed by a new stress accumulation and a new stress drop associated to a new slip. Tests were performed on intact rocks, already fractured rocks, different fracture's surfaces and fillings, and for each test, stick-slip was observed. The evolution of the friction in the stick-slip model can be explained by the rate-and-state friction law (Scholz, 1998):

$$\tau = \sigma'_n \left[\mu_0 + a \ln\left(\frac{V}{V_0}\right) + b \ln\left(\frac{\theta V_0}{D_c}\right) \right] \quad (1.2)$$

Where τ is the shear stress, σ'_n the effective normal stress, V is the slip velocity, V_0 a reference velocity, μ_0 is the steady state friction at $V = V_0$ and θ a state variable (refers to the surface properties). Parameters a and b are scalar constants depending on local conditions (Cornet, 2016) and represent the magnitude of change in the friction coefficient. Scholz (1998) observed that if $a - b > 0$ the system is *hardening* (increase of friction coefficient with increasing velocity in comparison to μ_0), while it is *weakening* if $a - b < 0$ (decrease of friction coefficient with increasing velocity in comparison to μ_0). These processes are called *velocity hardening* and *weakening*, and correspond to stable (aseismic) and unstable (seismic) slip, respectively. Taking a spring-block system as analogy, the stability of a slip depends on the effective normal stress, the shear stress, the spring's stiffness k (representing the elasticity of the material), parameters a and b and the slip distance. The limit between stable and unstable slip conditions is given by a critical value of effective normal stress σ_c (Scholz, 1998) defined by :

$$\sigma_c = \frac{kD_c}{-(a - b)} \quad (1.3)$$

For $\sigma'_n > \sigma_c$, the system is unstable and a seismic slip occurs, while it is stable if $\sigma'_n < \sigma_c$, meaning that a decreasing effective normal stress leads the fault closer to a stable slip, i.e., an aseismic movement. From the Coulomb failure criterion, we have seen that a decreasing effective normal stress affects fault's stability by decreasing the shear strength, leading the fault closer to rupture. From the rate-and-state friction law we have seen that, if slip occurs, (1) it can be stable or unstable, depending on whether the friction increases (hardening) or decreases (weakening), respectively and (2) a decreasing effective normal stress leads towards a stable slip ($\sigma'_n < \sigma_c$). Thus, a decreasing effective normal stress does not necessarily induce an earthquake in a critically stressed environment.

Hydromechanical coupling

The effective normal stress, and hence fault stability, is highly dependant on the pore pressure, itself controlled by the presence of fluids. Conversely, the presence of fluids in the earth's crust is dependant on rock properties (porosity and permeability) as well as on the presence of deformation structures such as fault zones. This interconnection between faults and fluids are grouped under the term of hydromechanical coupling. A fault zone can be considered as a conduit or a barrier to fluids (or a combination of both along the fault plane). Their role as a barrier or a conduit is dependant on their internal structure and composition as it is proposed

by Caine et al. (1996). A fault zone is commonly characterized by a fault core and a damage zone. In their review on recent developments concerning fault's zone properties, Faulkner et al. (2010) define the fault core composition as gouge, cataclasite or ultracataclasite and the damage zone as consisting of fractures varying in size and density. The compositions of both, the fault core and the damage zone, lead to spatial and temporal variation of permeabilities (Caine et al., 1996; Evans et al., 1997). Shear-slip of a fault may increase permeability in response to dilation. This has been observed in laboratory tests by Seront et al. (1998), as well as in hydraulic shearing experiments related to geothermal reservoir characterization (Håring et al., 2008; Evans, 2005). Changes in hydrogeologic conditions after large earthquakes, such as stream discharge or groundwater levels variations have also been reported (Manga et al., 2016; Kinoshita et al., 2015; Shi et al., 2015). Increasing groundwater temperature was observed following the 1999 Chi-Chi earthquake in Taiwan: simulation of the measured temperature evolution revealed that permeability was enhanced by two orders of magnitude for the model to fit the measured data (Wang et al., 2013). Elkhoury et al. (2006) analyzed the response of the water level in wells to Earth tides to measure the evolution of permeability over 20 years in southern California. They observed that after relatively important earthquakes, there was a phase shift response in the water level in wells to the Earth tide component, suggesting a change in permeability. Authors used the observed phase lag to measure permeability, and suggested an increase in permeability by a factor as high as three. Evidence for increasing permeability has also been observed following the 1989 Loma Prieta earthquake in California (Rojstaczer et al., 1995), where the stream flow of a nearby river increased rapidly following the earthquake (within 15 minutes after the earthquake) as well as the electrical conductivity of the water. Authors suggested, based on the increase of the stream flow, that permeability was enhanced by one order of magnitude. Variation in the chemical composition of groundwater following a M_L 5.1 earthquake was observed in eastern France, 20 km NE of Besançon, in the Jura Mountains (Charmoille et al., 2005), where the electrical conductivity increased significantly 7 hours after the earthquake. Authors suggested that this increase in electrical conductivity was the result of an enhancement of fracture permeability allowing the mixture of old mineralized groundwater with younger groundwater of the karst system. While these example show how fault rupture may influence hydraulic properties and hence fluid flow, the presence of fluid can also influence fault rupture as mentioned in section 1.2, where an increasing pore pressure can affect the stress state on a fault.

1.3 State of stress of the earth crust

As mentioned in Zoback et al. (1989), earthquakes are, together with global plate tectonics, geodynamic processes that are driven by tectonic stresses within the upper lithosphere. These stresses mainly result from plate-motion, flexure of the lithosphere due to loading or unloading and thermoelastic forces (e.g. cooling oceanic lithosphere). These broad-scaled forces are generally uniform over distances that are 2 to 3 times greater (sometimes even 100 times) than the thickness of the elastic lithosphere. In addition to these tectonic forces, local effects of topography, anisotropy of physical parameters (e.g. density, elasticity), as well as erosion or man-made excavations result in local stress variations, which are much more localized. The

study of [Zoback et al. \(1989\)](#) aims at illustrating the global tectonic stress pattern in the earth crust by compiling, in the frame of the World Stress Map project initiated in 1986 by Mary Lou Zoback, worldwide in situ stress measurement from a variety of geophysical and geological measurements. Results confirms the regional uniformity in the stress patterns, even though the measurement techniques and investigated depth are different. Moreover, this study reveals that the maximum horizontal stress component is sub-parallel to the absolute plate motion, which is confirmed by [Heidbach et al. \(2018\)](#), who took the mean orientation of the maximum horizontal stress (considering the 30 years of additional data) and found similar orientations between the maximum horizontal stress orientation and absolute plate motion. In addition the global stress pattern, worldwide in situ stress measurements indicate that the earth crust is critically stressed, as suggested for example by the widespread occurrence of induced seismicity. Other interesting conclusions can be drawn from these experiments: pore pressure up to great depth (>10 km) is near hydrostatic and the bulk permeability, inferred from hydraulic tests in boreholes or from induced seismicity, seems to lie between 10^{-17} to 10^{-16} m² over a 1 to 10 km scale. These conclusions are the heart of the paper from [Townend and Zoback \(2000\)](#), who suggest that the crust is in a failure equilibrium state and thus allows for faults to continuously fail, promoting high permeable pathways from the surface towards depth and hindering pore pressure to be much greater than hydrostatic. To support their conclusions, the authors refer to the results of [Barton et al. \(1995\)](#), who analyzed the relationship between the state of stress, fracture and fault orientation and permeability. This study shows how favourably oriented structures, i.e. oriented in a way that favors failure under a given state of stress, have a much higher permeability than the ones which are not. Thus, active faults zones, through the brecciation process during slip counteracting fault sealing processes (e.g. fracture cementation, see for example [Kettermann et al. \(2020\)](#)), can be considered as permeable pathways for fluids transportation over great depth ([Townend and Zoback, 2000](#); [Sibson et al., 1992](#)).

1.4 Objectives

The aim of this study is to investigate existing fault criticality using naturally induced earthquakes occurring in karstic regions following heavy rainfalls, as well as to contribute to a better understanding of seismic processes and their link with groundwater circulation in the study area. To this end, the objectives are to:

- 1) Identify the mechanisms leading to fault rupture following important precipitation and or snow melt periods and develop a conceptual model for rain-triggered seismicity;
- 2) Develop an analytical model for pore pressure variations resulting from the above mentioned mechanisms using simple 1D solutions;
- 3) Collect continuous karstic spring discharge and groundwater level measurement in the vicinity of fault zones in the Jura Mountains;
- 4) Monitor the microseismic activity of locally important strike-slip faults in the Jura Mountains;

- 5) Identify events and provide a local catalogue;
- 6) Determine relations between increasing spring discharge rates, groundwater table fluctuations and earthquakes occurrence for the study site and apply the analytical model to the collected data;
- 7) Generate a stress model of the shallow earth's crust based on field data.

1.5 Thesis structure

The structure of the thesis is built around the objectives listed above and consists in 4 main chapters. Chapter 2 presents the concept of rain-triggered seismicity in karstic regions and the existing case studies that have been reported in literature. We identify and present the different mechanisms responsible for rain-triggered seismicity and apply simple 1D analytical solutions to the existing case studies to evaluate which mechanisms is more likely to trigger seismicity following intense precipitations. Chapter 2 is planned for publication and is thus written as a stand-alone manuscript. Chapter 3 consists mainly in the presentation of the collected data: the karstic spring hydrographs and the seismic catalog. In this chapter, we also perform a preliminary analysis of the relation between an increasing hydraulic gradient and earthquake occurrence and determine which earthquakes are potentially rain-triggered. In Chapter 4, we make a detailed analysis of the seismicity of the different fault zones and apply our analytical model to our data for events identified as potentially rain-triggered. For each fault zone showing seismicity following heavy rainfalls we define a triggering pore pressure responsible for earthquake occurrence and discuss the results with regards to the geological and hydrogeological context of the study area. In Chapter 5, we derive a Best Estimate Stress Model from our data and evaluate how our findings fit in the general tectonic context of the Jura Mountains. Finally, in Chapter 6, we synthesize the major outcomes of each chapter and give perspectives for future research.

Chapter 2

Rain-induced seismicity in karstic regions

2.1 Introduction

Water at elevated pressures plays a key role in earthquake mechanics. Direct observation of earthquake triggering related to water includes (1) water injection in deep wells (Parotidis et al., 2004; Evans et al., 2012; Keranen et al., 2013; Ellsworth, 2013; Diehl et al., 2017), (2) reservoir impoundment (Kangi and Heidari, 2008; Gupta, 2002) and (3) seasonality or cyclicity in the occurrence of earthquakes due to natural phenomenon (Saar and Manga, 2003; Scholz et al., 2019). The latter includes case studies having reported seasonal patterns of seismicity and their relation to groundwater recharge or snow load (Wolf et al., 1997; Saar and Manga, 2003; Bettinelli et al., 2008; Braunmiller et al., 2014; Birhanu et al., 2018; Bragato, 2021). Triggering is observed in a variety of tectonic setting, from intraplate regions to volcanic islands. These regions have accordingly different hydrogeologic settings and climate conditions, and show a broad range of time lag between the rainfall and the seismicity, varying from hours to months (Costain and Bollinger, 2010). Three case studies show relatively small time lags (hours to days), and have similar local geology. As pointed out by Miller (2008), the cases of Hochstaufen (Germany) (Kraft et al., 2006b), Muotathal (Switzerland) (Husen et al., 2007) and Avignon region (France) (Rigo et al., 2008) all occur in known karstic regions. The primary mechanism invoked for explaining seismicity following rainfall is pore fluid pressure diffusion, similarly to what is described in reservoir induced seismicity (Talwani, 1997; Talwani and Acree, 1984), considering a homogeneously fractured crust. However, for the cases of Germany, Switzerland and France and considering the intrinsic properties of karst, the model of a homogeneously fractured crust might be too simplistic, leading to the necessity to explore other possible mechanisms.

The study aims at identifying and differentiating what mechanism is more likely to induce seismicity following heavy rainfalls according to local stress conditions and local geology by quantifying their effects using simplified analytical solutions on data from the above-mentioned case studies (Germany, Switzerland and France). The analytical modeling requires several values to be estimated, based either on average values from literature or using data of existing case studies. We present one scenario for each case study to illustrate orders of magnitude between the different mechanisms for a given set of conditions. The aim is to compare the mechanisms with

each other, not to find the best fitting parameters nor to propose an extensive and systematic sensitivity analyses.

2.2 Karst geology, triggering mechanisms and conceptual model

Karst specific features

Mature karst systems consist in well developed channels of various morphologies within a relatively low permeable rock matrix, typically consisting of soluble rocks such as limestones, dolomites or evaporites. Different hydraulic parameters between channels and matrix lead to the so-called duality of karst (Király, 1994, 1998). On the surface, the epikarst, consisting in a highly altered surface bedrock, is thought to enhance the concentrated infiltration within the karst conduits (Király, 1998). Bonacci (1987) carried out numerous analyses on karst and found that the effective porosity of karst host rock is between 0.1 to 1% on average, being locally much higher in presence of large caves or conduits. This configuration, i.e. high permeable conduits and low porosity matrix, satisfactorily explains karstic spring hydrographs following recharge events. Figure 2.1 illustrates the characteristic hydrographs of springs related to a) a fractured rock mass b) a bare karst and c) a fractured rock mass with karst conduits. The shape of hydrographs depends on the extend of the fractures within the host rock. The hydrograph resulting from the combination of a fractured host rock with karst conduits (Figure 2.1c) represents the global response of both the high permeable conduits and low permeable matrix in response to recharge events. In this scenario, the recharge event involves inversion of the flow regime. During flood events, the evacuation capacity of the karst system is exceeded and results in rapid and large water levels increase within the channels. This leads to a flow from the channels to the matrix, recharging the matrix. During low-water periods, the water stored in the matrix flows back to the conduits, maintaining base flow levels. Such hydrographs are typical of karstic regions and are clear evidence of rapid hydraulic head increases. For example, Figure 2.2 shows the discharge of the Schlichenden Spring draining the Hoelloch cave system (Muotathal, Switzerland). During an important recharge event, a cave located 540 m above the permanent water table was flooded (Miller, 2008). Another important aspect to consider is the relationship that can exist between karst system and fault zones. A fault zones's hydraulic properties depend on various parameters and are subject to changes over time and space. Considered either as barrier or conduits for fluid circulation, their role depends greatly on their anatomy (fault core and damaged zone) and composition (breccia, cataclasite or ultracataclasite) (Caine et al., 1996; Faulkner et al., 2010), which will influence the spatial distribution of permeabilities (Caine et al., 1996; Evans et al., 1997). The hydraulic properties of a fault zone can also vary in time as it has been observed in laboratory tests (Seront et al., 1998) and modelling experiments (Preisig et al., 2016), where shear slip on a fault increases the permeability in response to dilation. Since karst development is driven by water flow and faults can favor flow paths, it is likely that karst system develop in relation with faults and that the two system are hydraulically connected. However, the exact role of the fault, i.e. whether fluids can pass through the fault zone or flow along is determined by its internal structure and composition (Matonti et al., 2012). This potential

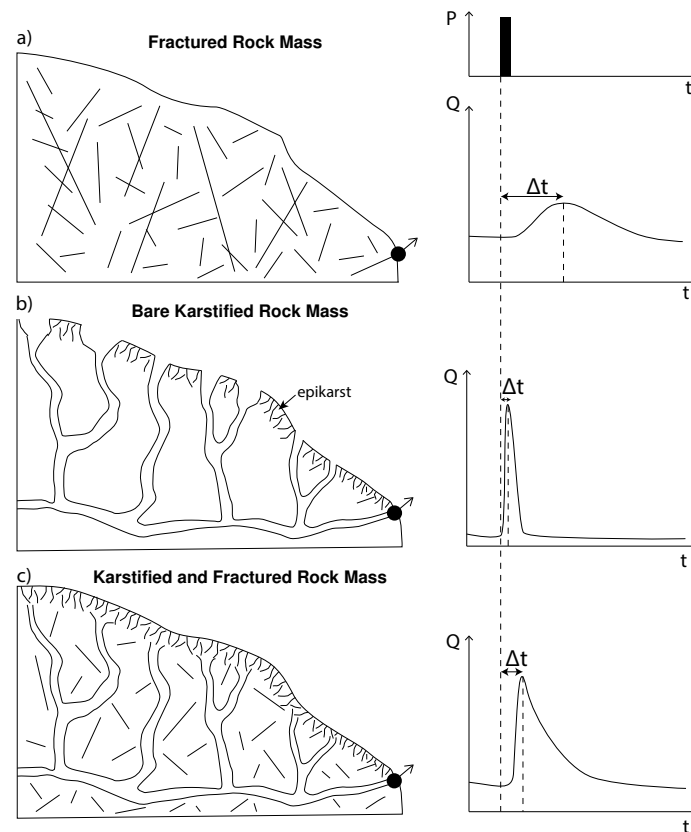


Figure 2.1: Characteristic spring hydrograph for a) a fractured rock mass, b) bare karst and c) karst conduits in a fractured host rock, for a same precipitation event (P). The time lag (Δt) between the maximum rain intensity and the maximum flowrate (Q) depends on the level of karstification. For a) (no karst) the time lag is longer than for b) (bare karst, short time lag) and c) (combined fractured and karstified rock mass, mixed response). Figure modified after Bonacci (1993).

link between karst and faults make the later particularly sensitive to rainfall and karst flooding events.

Possible effects of rainfall on fault's stability

External changes in the stress regime might impact fault's stability depending on its orientation, initial stress regime and pore pressure. We identify four potential mechanisms to explain the influence of rainfall on fault stability:

- i. crustal loading,
- ii. poro-elastic pore pressure changes associated with crustal loading,
- iii. pore pressure diffusion and,
- iv. direct hydraulic connection

The impact of crustal loading (i), i.e. the weight of precipitation as an additional load on the crust, depends on the stress regime. From Anderson's stress regime classification (Anderson, 1951), an additional load in a normal faulting stress regime (NF) reduces fault stability, increases

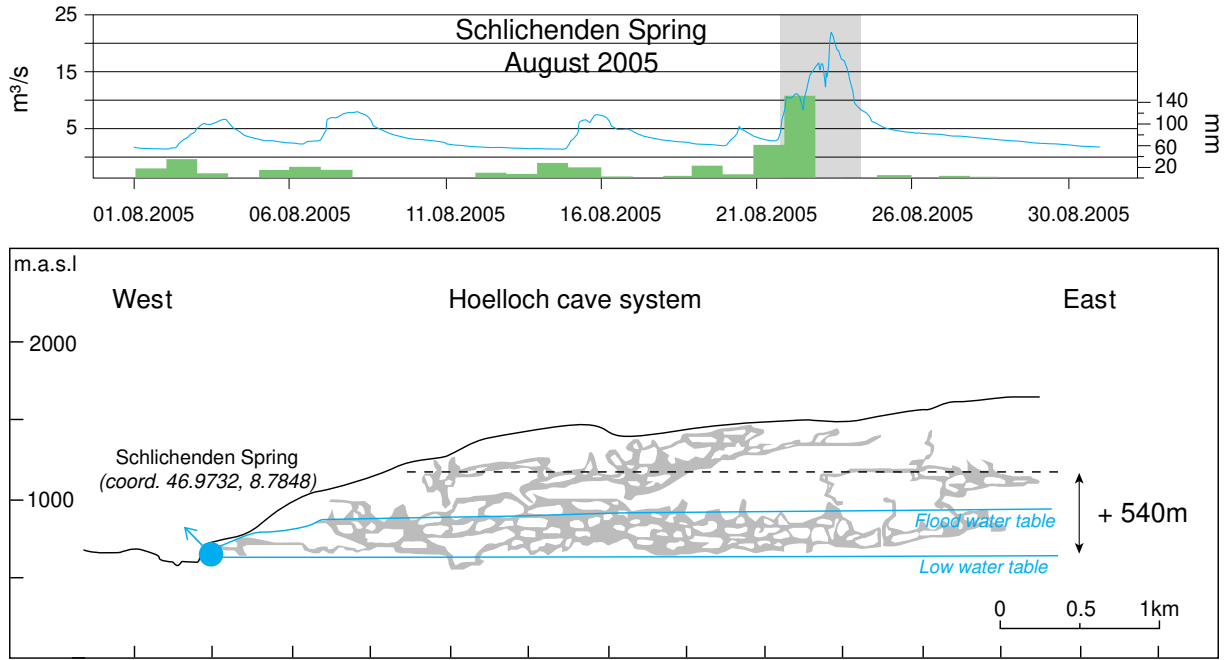


Figure 2.2: Top: Discharge curve (blue line [m^3/s]) at the Schlichenden Spring during the month of August 2005 (Federal Office for the Environment - Hydrological data and forecast) and the daily precipitations (green bars [mm]) (Federal Office of Meteorology and Climatology MétéoSuisse). The shaded area illustrates the time at which a cave located at +540m from the spring altitude was flooded. Bottom: Cross section through the Hoelloch cave system with the low and flood water tables (blue) and the level at which the cave was flooded (dashed line) in August 2005. Figure modified after Wildberger et al. (2010).

stability in a thrust faulting stress regime (TF) and leaves the stability unchanged in a strike-slip stress regime (SS). The additional load on the crust also modifies normal effective stresses due to increasing pore-fluid pressure through poro-elastic deformation (ii), reducing the shear strength and hence fault stability. A rising fluid pressure at the surface might also initiate a pressure diffusion front (iii) increasing pore-pressure and reducing the effective normal stress gradually while the pressure front propagates. A rising water level and a direct hydrogeological connection through the karst between the surface and the fault plane (iv), results in an instantaneous increase in pore pressure impacting fault's stability.

Conceptual model

We propose a conceptual model (Figure 2.3) that considers specific karst features and rainfall induced mechanisms that potentially lead to fault instability. An important difference between these mechanisms depends on the relation between the karstic network and the faulted system, whether a direct hydraulic connection exist (iv) or not (ii and iii). In cases where the karstic network is connected to a highly permeable fault zone (iv), the groundwater pressure increase at a given depth is quasi-instantaneous. In cases where the karstic network lies above a fault, and there is no direct hydraulic connection, the mechanism leading to stress changes is similar to the one described for reservoir impoundment, that is (ii) a rapid increase in pore pressure resulting from the elastic deformation due to the vertical load and undrained conditions and/or (iii) a delayed response due to fluid migration towards depth. Although an increasing

groundwater pressure occurs in all scenarios, they can be differentiated by the timing, which is quasi-instantaneous for (ii) and (iv) and delayed for (iii).

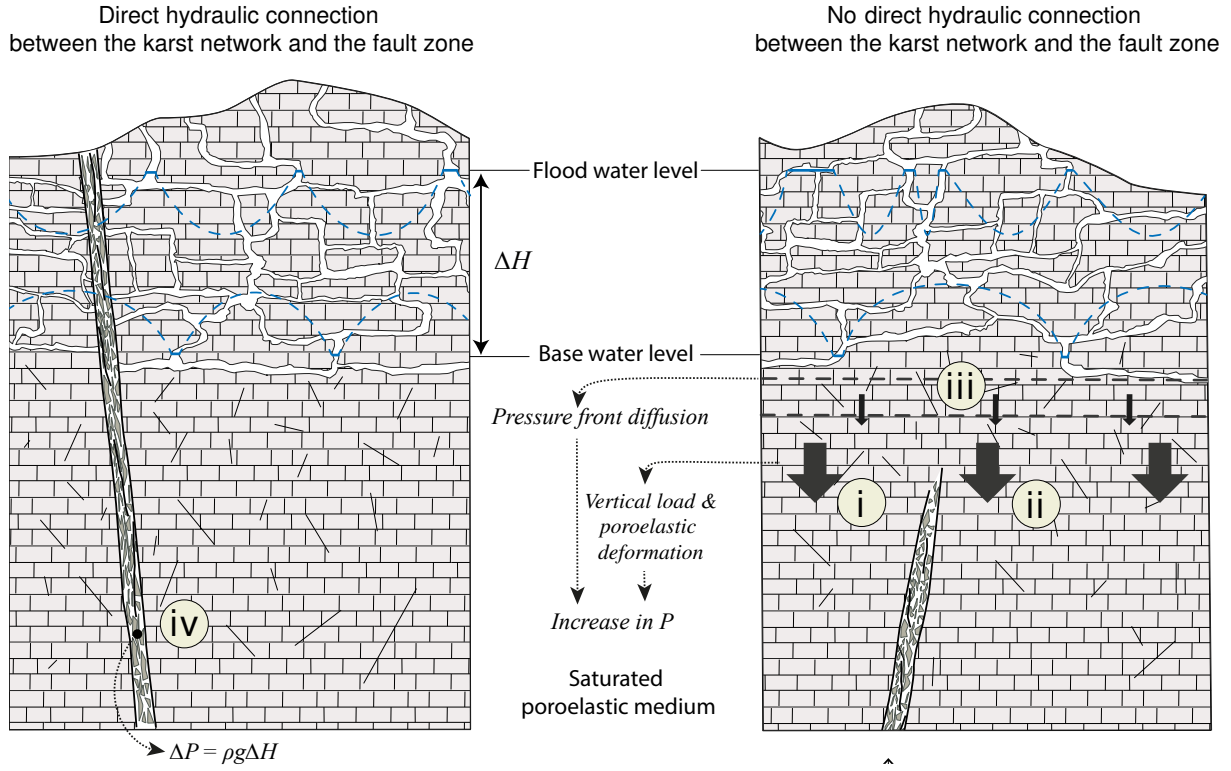


Figure 2.3: Schematized scenarios illustrating mechanisms leading to pore pressure (P) increase (ΔP) following an increasing hydraulic head (ΔH) in the case where the karstic network is in direct hydraulic connection with a fault (left), or lying above (right). The vertical load (i) increases pore pressure through poro-elastic deformation (ii). (iii) Pore pressure increase is delayed due to the pressure front migration. (iv) Pore pressure increases quasi-instantaneously following an increasing water level. Note that (i), (ii), (iii) and (iv) might act concomitantly.

2.3 Methodology

Fault stability and Coulomb failure stress

According to the classical Mohr-Coulomb failure criterion, failure on a fault initiates when the following stress conditions are met:

$$\tau = c + \mu \sigma'_n \quad (2.1)$$

where τ is the shear stress, μ the coefficient of internal friction, the effective normal stress $\sigma'_n = \sigma_n - p$, with σ_n the normal stress, c the cohesion and p the pore pressure in the fault zone. The coefficient of sliding friction μ typically lies between 0.6 and 0.85 (Byerlee, 1978) and in the context of reactivation, faults are often considered to be cohesionless ($c = 0$) (Sibson, 1985). For constant fault properties, the external parameters affecting fault stability are the shear stress τ and the effective normal stress σ'_n , which are both dependent on the fault's orientation (angle θ between σ'_1 and the normal to the fault plane) and stress magnitude as expressed in a 2D plane

strain conditions or when the intermediate stress is co-planar with the fault by the following relations (Jaeger et al., 2009):

$$\sigma'_n = \frac{\sigma'_1 + \sigma'_3}{2} + \frac{\sigma'_1 - \sigma'_3}{2} \cos(2\theta) \quad (2.2)$$

$$\tau = \left| \frac{\sigma'_1 - \sigma'_3}{2} \sin(2\theta) \right| \quad (2.3)$$

For a fault favourably oriented ($\theta = 60^\circ$) and according to Eq. (2.1), (2.2) and (2.3) rupture can be initiated by an increasing shear stress τ through an increase of σ'_1 and/or a decrease in σ'_3 or through a decrease of the effective normal stress σ'_n .

To evaluate the proximity of a fault to failure, the Coulomb failure stress (CFS) is a good indicator and is defined as follows (see e.g. Beeler et al., 2000):

$$CFS = \tau - \mu(\sigma_n - p) - c \quad (2.4)$$

A negative value resulting from Eq. (2.4) indicates that failure has not been reached, as the shear strength ($c + \mu\sigma'_n$) is greater than the shear stress (τ). The change in Coulomb failure stress shows the impact of stress changes on the fault's stability. Coulomb failure stress change is defined as follows (Beeler et al., 2000):

$$\Delta CFS = \Delta\tau - \mu(\Delta\sigma_n - \Delta p) \quad (2.5)$$

where $\Delta\sigma_n$ the change in normal stress, is positive for compression. Positive values of ΔCFS indicate that the fault is closer to failure, whereas negative values indicate that the fault is moving away from the rupture point.

Direct connection

If the fault zone is in direct hydraulic connection with the karstic network, and considering a simplified environment with no pressure losses, an increasing hydraulic head in the karst conduits due to heavy precipitation will generate a quasi-instantaneous and non-attenuated pressure signal. The increase in pore pressure (and the decrease of the effective principal stresses σ'_1 and σ'_3) will directly relate to the increase in the hydraulic head, Δh , by the relation $\Delta P = \rho g \Delta h$, leading to a decreasing σ'_n . The variation in the hydraulic head Δh [m] depends on the storage coefficient S [-], a considered area A [m²] and the volume of infiltrated water V_w [m³], which is controlled by the (effective) rainfall intensity i [m/day] falling over the area A during a given time dt [day] (Fetter, 2001):

$$\Delta h = \frac{V_w}{S A} \quad (2.6)$$

where

$$V_w = i dt A \quad (2.7)$$

Combining Eqs (2.6) and (2.7) gives:

$$\Delta h = \frac{i}{S} \frac{dt}{dt} \quad (2.8)$$

In a porous rock mass in which the water table can fluctuate freely, S is equal to the effective porosity, which can range from less than 0.01 to 0.30 [-] in sedimentary karstic rocks. In a porous rock mass in which the water table cannot fluctuate freely, i.e. the piezometric level is above the top of the permeable unit (confined aquifer), S is in the order of 0.005 [-] or less (Fetter, 2001).

Crustal loading

Considering a karst network without a direct hydraulic connection with the fault zone, fault stability might be influenced by a supplementary load, i.e. the infiltrated water. In such a case, an underlying saturated fault system will behave as a poro-elastic medium with increasing pore pressure (and decreasing effective stress) in response to the vertical loading.

Boussinesq's solution

To evaluate the impact of an increasing load P_a on the vertical stress $\Delta\sigma_v$ beneath the center of a karstic network, the Boussinesq's solution for a circular load on the surface of a semi-infinite elastic space is used, as expressed in Poulos and Davis (1974):

$$\Delta\sigma_z = P_a \left[1 - \frac{1}{[(R/z)^2 + 1]^{3/2}} \right] \quad (2.9)$$

where R is the radius of the circular area, and can be seen as the catchment area of the karstic network and z is the depth. This solution is then used to calculate the change in pore-pressure resulting from the change in vertical stress (Eq. (2.11)).

Poro-elastic deformation

Under undrained conditions, the impact of stress changes on pore pressure is expressed as follows (Rice and Cleary, 1976):

$$\Delta p = B \frac{\Delta\sigma_{kk}}{3} \quad (2.10)$$

where $\Delta\sigma_{kk}/3$ corresponds to $(\Delta\sigma_x + \Delta\sigma_y + \Delta\sigma_z)/3$ and is the average compressive stress. Assuming that stress changes in only the vertical direction, then Eq. (2.10) becomes :

$$\Delta p = \frac{B(1 + \nu_u)}{3(1 - \nu_u)} \Delta\sigma_z \quad (2.11)$$

where B is the Skempton coefficient and ν_u the undrained Poisson's ratio.

Fluid pressure diffusion

An increasing hydraulic head will lead to fluid pressure diffusion from zones of high to zones of low hydraulic head. Considering a step increase in pressure at time $t = 0$, the simplified 1-D diffusion problem has the solution (Carslaw and Jaeger, 1959):

$$p(z, t) = P_a \operatorname{erfc} \left(\frac{z}{2\sqrt{Dt}} \right) \quad (2.12)$$

where erfc is the complementary error function, p is the pressure at depth z and time t resulting from the applied pressure P_a on a medium with diffusivity D , where $D = T/S$, with T the transmissivity and S the storage coefficient.

Now considering a step increase in pressure at time $t = 0$ and a linear decrease over a time τ from P_a to 0, the solution reads as follows (Perrochet P., personal communication, presented in Appendix A) :

$$p(z, t) = P_a \operatorname{erfc} \left(\frac{z}{2\sqrt{Dt}} \right) - m \left\{ \left(t + \frac{z^2}{2D} \right) \operatorname{erfc} \left(\frac{z}{2\sqrt{Dt}} \right) - \frac{z\sqrt{t}}{\sqrt{\pi D}} \exp \left(\frac{-z^2}{4Dt} \right) \right\} \\ + H(t - \tau) m \left\{ \left(t - \tau + \frac{z^2}{2D} \right) \operatorname{erfc} \left(\frac{z}{2\sqrt{D(t - \tau)}} \right) - \frac{z\sqrt{t - \tau}}{\sqrt{\pi D}} \exp \left(\frac{-z^2}{4D(t - \tau)} \right) \right\} \quad (2.13)$$

where $H(u)$ is the Heaviside step function ($H(u) = 1$ when $u > 0$ and $H(u) = 0$ when $u < 0$), and $m = P_a/\tau$. Note that Eqs (2.12) and (2.13) do not consider the coupled poro-elastic response, the rock and pore space deformation associated with pressure changes.

2.4 Case studies

Three different case studies where rain-induced seismicity occurred in karstic regions after a period of intense rainfall are presented and discussed: (1) Mount Hochstaufen in Germany (Kraft et al., 2006b; Hainzl et al., 2006), (2) Muotathal and Riemenstalden in Switzerland (Husen et al., 2007) and (3) the Avignon region in France (Rigo et al., 2008). In the following, we present these case studies in a consolidated and homogeneous manner to facilitate comparisons. Relevant data of the three case studies will be considered as input for our simplified analytical models.

Mount Hochstaufen, Germany

In 2001, the University of Munich, together with the Bavarian State, installed a new seismic monitoring network in Bavaria, as well as a subnet of 6 stations around Mount Hochstaufen (Kraft et al., 2006b). These were installed following preliminary observations made by Schwarzmann (1996), who analysed 2 months of seismic records from a temporary network installed after a M_L 2.7 on June 26th 1995 in the region of Bad Reichenhall. The author suggested a possible relation between seismicity and rainfall.

Mount Hochstaufen experienced two seismic swarms in 2002, both occurring after an extended period of rainfall in March and August. In August, ~ 300 mm fell over the region of Bad Reichenhall (German Weather Service, DWD) in less than one week, with a maximum

intensity of ~ 120 mm/day and ~ 110 mm/day on the 6th and the 11th of August, respectively. In all, 1171 earthquakes were recorded with the maximum event magnitude of M_L 2.4. The located events have epicentres distributed over the Staufen Massif with focal depth mostly lying between 0 and 3 km depth. Figure 2.4 summarizes the geological context around the Staufen Massif, which is part of the Northern Calcareous Alps (NCA), forming an east-west striking fold-and-thrust belt. The NCA are part of the Upper Austroalpine unit and are commonly divided into three main nappes unit : (1) Bajuvaric, (2) Tirolic and (3) Juvavic. Mount Hochstaufen is part of the Tirolic nappes overthrusting the Bajuvaric Nappes, which are overthrusting the Helvetic and Penninic units (Figure 2.4a and b) (Henrich and Zankl, 1981; Tollmann, 1976). Mount Hochstaufen is located south of this major thrust fault oriented E-W and north of an important SW-NE sinistral strike-slip fault called ISAM fault (Innsbruck-Salzburg-Amstetten). Contemporaneous stress field measurement from borehole breakouts illustrated on the Stress Map Germany (Reiter et al., 2016) suggest a N-S maximum horizontal stress orientation.

The stratigraphic sequence of the Staufen Massif consists in units from the lower to middle Triassic with lithologies mainly dominated by limestones and dolomites. Figure 2.4c and 2.4d show the lithologies, along with their hydraulic properties (Bayerisches Landesamt für Umwelt, last accessed in January 2021). The higher part of the Staufen Massif are constituted of Wetterstein limestone (Tollmann, 1976). This formation, along with the Dachstein formation outcropping in the southern and eastern NCA, is the most important karst host rock in the NCA.

For the analytical modelling, we consider an average altitude of 1500 m.a.s.l and an average focal depth of 1000 m.b.s.l.. The altitude of a perennial karstic spring located 4.5 km west of Hochstaufen is taken as base water level. The horizontal distribution of epicenters simplified as a circle with a radius of 5 km is also considered.

Muotathal and Riemenstalden, Switzerland

In August 2005, the regions of Muotathal and Riemenstalden in central Switzerland experienced a succession of events over a 12-hour period after four days of heavy rain with a total of ~ 200 mm of precipitations, with maximum intensity of ~ 75 and 90 mm/day on the 21 and 22 of August. The return period of such intensive rainfall periods is approximately 300 years (Husen et al., 2007). A total number of 47 earthquakes were recorded and distributed in two clusters (regions of Muotathal and Riemenstalden). The focal depths are between 0 and 5 km. Muotathal and Riemenstalden are both located in the Helvetic unit in central Switzerland and are part of the Axen nappe (Pfiffner, 2011). The Axen nappe consists of formation from the Triassic to the Paleogene. In the area of interest, the lithologies are dominated by Cretaceous limestones. Due to extensive thrusting, the units are stacked on top of each other, leading to an important thickness of Cretaceous units (Schmid et al., 2004).

A major thrust delimits the Axen nappe with the Drusberg nappe close to the villages of Muotathal and Riemenstalden. Numerous fault lineament are visible on the geological map and show a N-S general trend as well as minor lineament showing an WNW-ESE strike. Some of these faults show distinct horizontal offsets indicating a strike-slip environment. Data provided

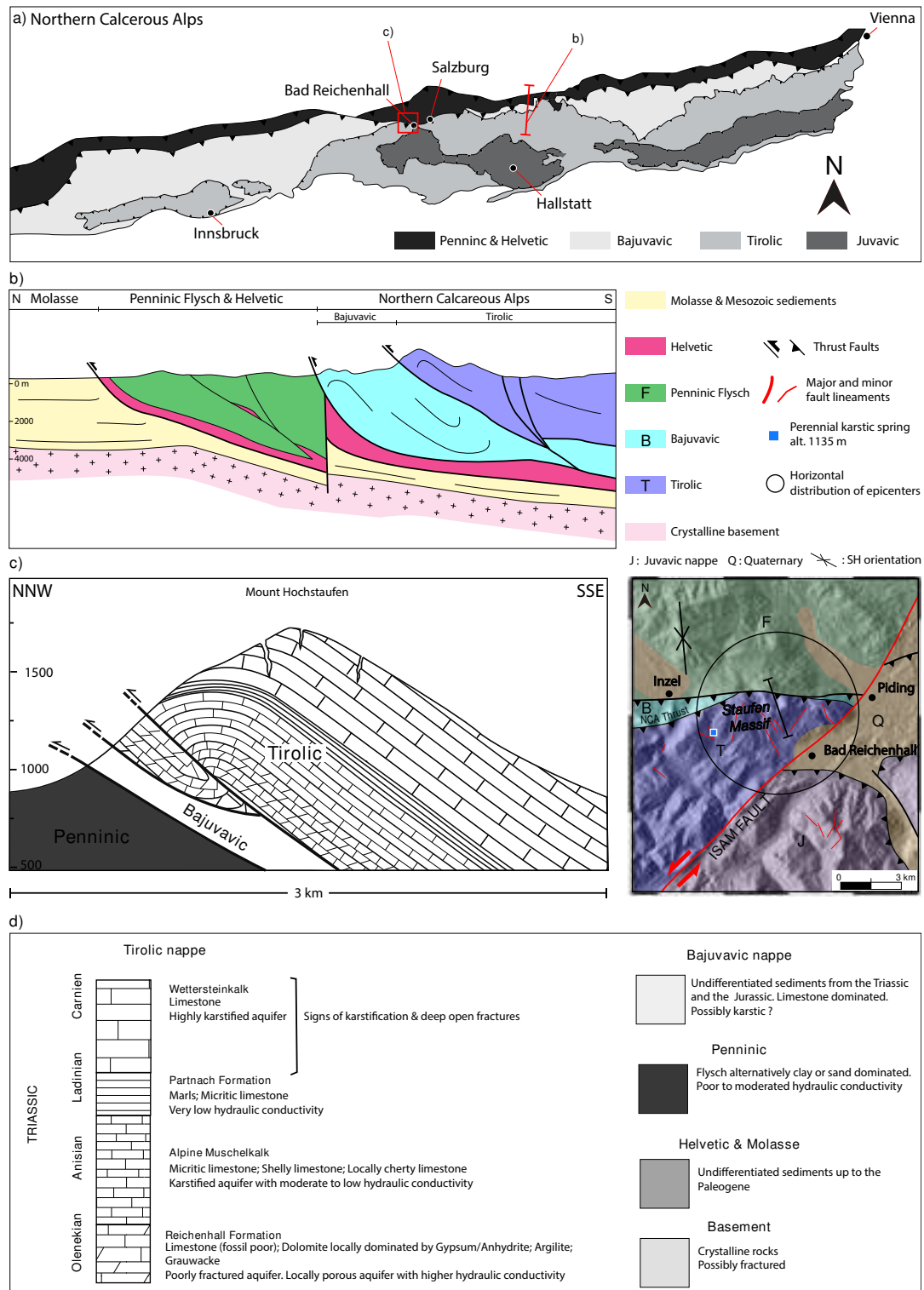


Figure 2.4: a) Northern Calcareous Alps (NCA) with the different nappes and indication for locations of b) and c). b) Simplified cross section through the northern part of the NCA illustrating the successive thrusting, modified after [Leitner and Spötl \(2017\)](#). c) Schematic sketch of Mount Hochstaufen illustrating the different lithologies of the Tirolic unit (left) and simplified tectonic map around Mount Hochstaufen showing major deformation structures as well as the orientation of the maximum horizontal stress (SH), modified after [Schweigl and Neubauer \(1997\)](#), and d) main hydraulic properties of lithologies of the Tirolic nappe (left) as well as a brief description of the succession of the unit below the Tirolic towards the basement (right) ([Leitner and Spötl, 2017](#)).

by the World Stress Map (Reiter et al., 2016) suggest a $\sim$ SE-NW orientation of the maximum horizontal stress for this region (data inferred from focal mechanisms).

With a total length of more than 250 km of galleries (Wildberger et al., 2010), the region is a well known karstic area. The Hölloch cave and the Silberen system are part of this karstic network and are located east of the village of Muotathal. Karst developed mostly in the upper Cretaceous, in the Urgonian and Seewen limestones, and due to thrusting, the thickness of these karstified rocks can be doubled or tripled (Figure 2.5) (Wildberger et al., 2010). The lithologies and their hydraulic properties are presented in Figure 2.5d.

For the analytical modelling, we consider an average altitude of 1400 m.a.s.l. and an average focal depth of 2000 m.b.s.l.. The altitude of the perennial Schlichenden karstic spring, the outlet of the Hölloch cave system, is taken as base water level. The horizontal distribution of epicenters simplified as a circle with a radius of 2.5 km is also considered.

Avignon region, France

In south-eastern France, in the Gard department, a catastrophic storm occurred in September 2002. In two days, accumulated precipitation reached $\sim$ 400 mm in Pujaut Basin, near Avignon, and even reached 650-700 mm further to the north, near the city of Alès. Seismicity increased during the two weeks following the storm. 11 earthquakes were recorded within a temporary network centered around the Pujaut Basin, and 2 events were located outside the network, but still relatively close in distance. The focal depth are mostly lying between 0 and 10 km (Rigo et al., 2008).

The Gard department located in the Provence, lies between the Alps in the northeast and the Pyrenees in the southwest. The regional geology consists of a sedimentary cover of Mesozoic and Cenozoic ages, varying in thickness from 2 to 7 km (Rigo et al., 2008). Three major faults exists in the area of interest, the Nîmes Fault, the Pujaut fault and the Roquemaure fault. The Nîmes fault has a NE-SW orientation and has a senestral strike-slip movement (Schlupp et al., 2001). The Pujaut fault, parallel to the Nîmes fault, corresponds to a normal fault steeply dipping towards the north and is the result of the Oligocene extension regime (Jolivet and Faccenna, 2000). The northern compartment of the Pujaut fault result in a depression called the Pujaut basin or graben. The Roquemaure fault is considered as inactive or even nonexistent (Rigo et al., 2008). Further east, the Salon-Cavaillon fault system strikes SE-NW, has a normal component possibly combined with a dextral sense of movement (Terrier et al., 2007). From the reconstruction of large scale tectonic movements (Jolivet and Faccenna, 2000), we estimate the present-day maximum horizontal stress orientation to be approximately N-S to NW-SE.

The area is well known for its deep karstic conduits that developed during the Messinian crisis (Julian and Nicod, 1984; Audra et al., 2004). Exploration by a Remotely Operated Vehicle (ROV) in the Fontaine-de-Vaucluse brought evidence of former epiphreatic karst conduits down to -100m below sea level (Bayle and Graillot, 1987). From seismic data, drilling and field observations, evidence show that canyons are lying around 1000 m below sea level near the city of Avignon (Clauzon, 1982; Gargani, 2004). At the end of the Messinian crisis, the canyons were flooded and filled with terrigenous sediments (Clauzon, 1982). Figure 2.6c shows the

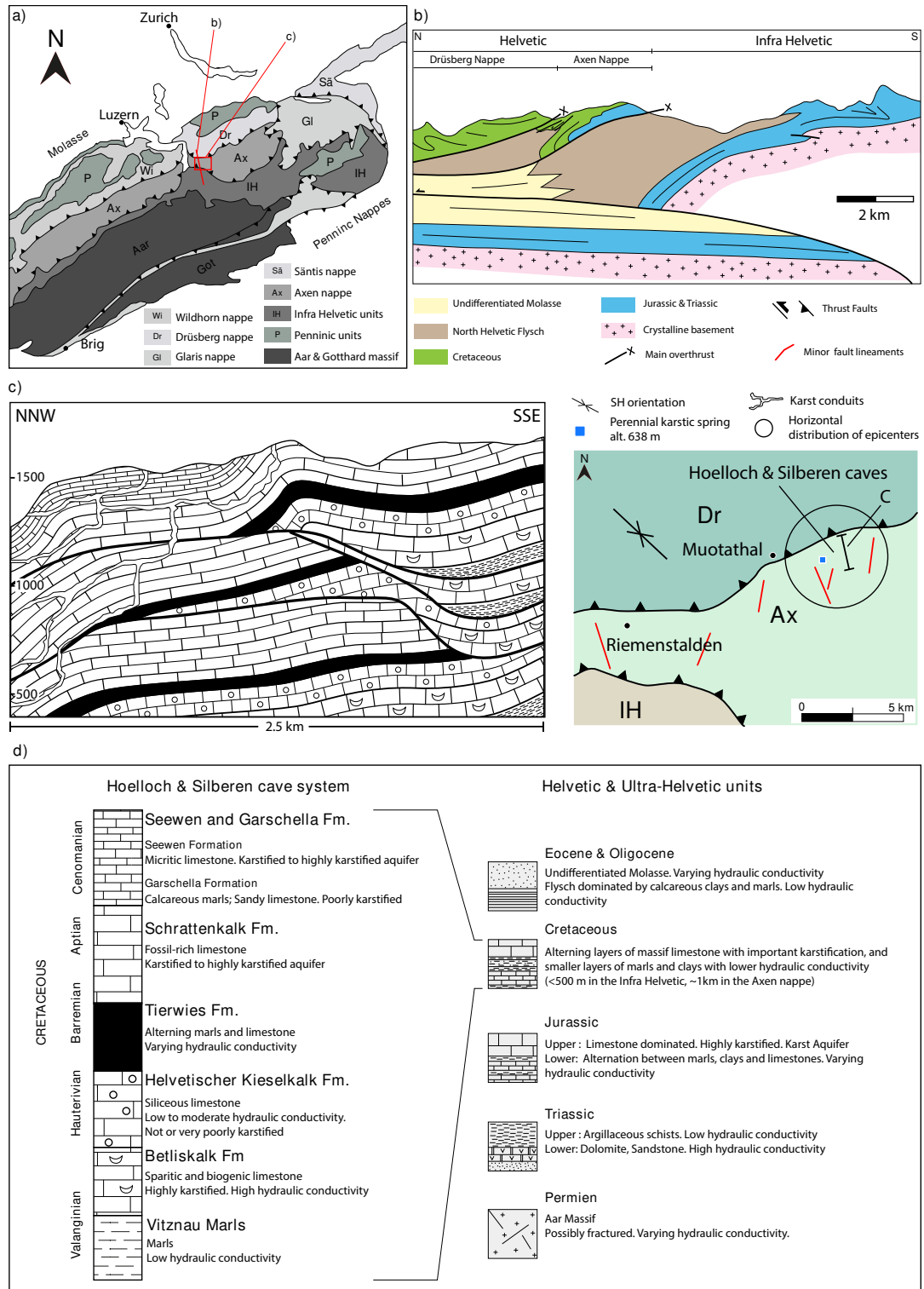


Figure 2.5: a) Tectonic map around central Switzerland and the different nappes and indications for b) and c). b) Cross section through the Helvetic and the Infra Helvetic near the locations of Muotathal and Riemenstalden, modified after Pfiffner (2011). c) (left) Schematic figure of the Hoelloch and Silberen cave system and (right) a simplified tectonic map around Muotathal and Riemenstalden (modified from the Swiss Tectonic Map, Swiss Federal Office for Topography, swisstopo) showing major deformation structures as well as the orientation of the maximum horizontal stress (SH). d) Lithologies and associated hydraulic properties of the different Helvetic and Ultra Helvetic units (Nagra, 1996).

relation between the deep karstic system and the ancient canyons and Figure 2.6d illustrates the lithologies in the area of interest.

For the analytical modelling, we consider an average altitude of 200 m.a.s.l. and an average focal depth of 3000 m.b.s.l.. The altitude of the Rhone River near Avignon is taken as base water level. The horizontal distribution of epicenters simplified as a circle with a radius of 19 km is also considered.

Analytical modelling

We quantify the effect of each mechanism on fault stability for the different case studies by considering a semi-infinite environment with initial stress conditions of σ_1 , σ_3 and the initial pore pressure p_{init} at depth z , with cohesion c and internal friction coefficient μ . The environment is saturated at depth and unsaturated towards the surface with a unsaturated layer of thickness z_{unsat} (see Figure 2.7). Further considered is a circular area on the surface of the saturated layer of radius r being loaded with pressure P , which is inferred from the rain intensity i . Values for z , z_{unsat} , r and P are based on the case studies, initial stress conditions are estimated using average lithostatic and hydrostatic pressure gradients of 0.0225 MPa/m and 0.0098 MPa/m, respectively. The depth z is inferred from focal depth, the unsaturated thickness z_{unsat} is calculated by subtracting the altitude of local springs from the average altitude of the area. The radius r corresponds roughly to the horizontal spatial distribution of seismicity. The pressure P is inferred from the rain intensity. In the case of a direct hydrogeologic connection, the pressure is quasi equivalent to ΔH (Eq. (2.8)). For the case there is no hydrogeologic connection the pressure load corresponds to the additional weight of infiltrated water. For the stress conditions, we consider a strike-slip environment where $\sigma_H = \sigma_1$, which is at least equal to σ_v , and $\sigma_h = \sigma_3$ which is set to the minimum possible value to have a $CFS < 0$, meaning that rupture has not yet occurred. Cohesion c is thought to be negligible and a value of 0.6 for the internal friction coefficient μ is taken. Input values for the analytical modelling are shown in Table 2.1. Precipitations are inferred from the case studies as shown in Figure 2.9. Note that we slightly simplify the rainfalls for the analytical modelling by considering only 2 rainfall events for each case study (i_1 and i_2). The storage coefficient S is set according to observed water level fluctuations in karstic network reported in literature, which ranges between a few tens of meters (Lesparre et al., 2017) to several hundreds of meters (Jeannin, 2001). For Muotathal, a water level increase of 540 m for 200 mm of rain corresponds to a storage coefficient of 0.0004 (Eq. (2.8)). Since Muotathal is the only case study where this value can be calculated, we decide to use a more conservative value of $S = 0.001$, which corresponds to a level fluctuation of 100m for 100mm of infiltrated water. Using Eqs (2.1)-(2.12) and data from Table 2.1, changes in pore pressure are calculated, as well as changes in the Coulomb Failure Stress (CFS). Results are separated between direct and delayed effects. Direct effects include changes in pore pressure resulting from the hydraulic head increase, as well as poro-elastic deformation resulting from an increasing vertical stress, while the delayed effect is a pore pressure increase resulting from the diffusion process. For poro-elastic computations, we assume undrained conditions. We consider high rainfall intensity events which implies that the drainage capacity of the karstic network is overcome during the rainfall. This result in a rapid increase in the hydraulic head within the

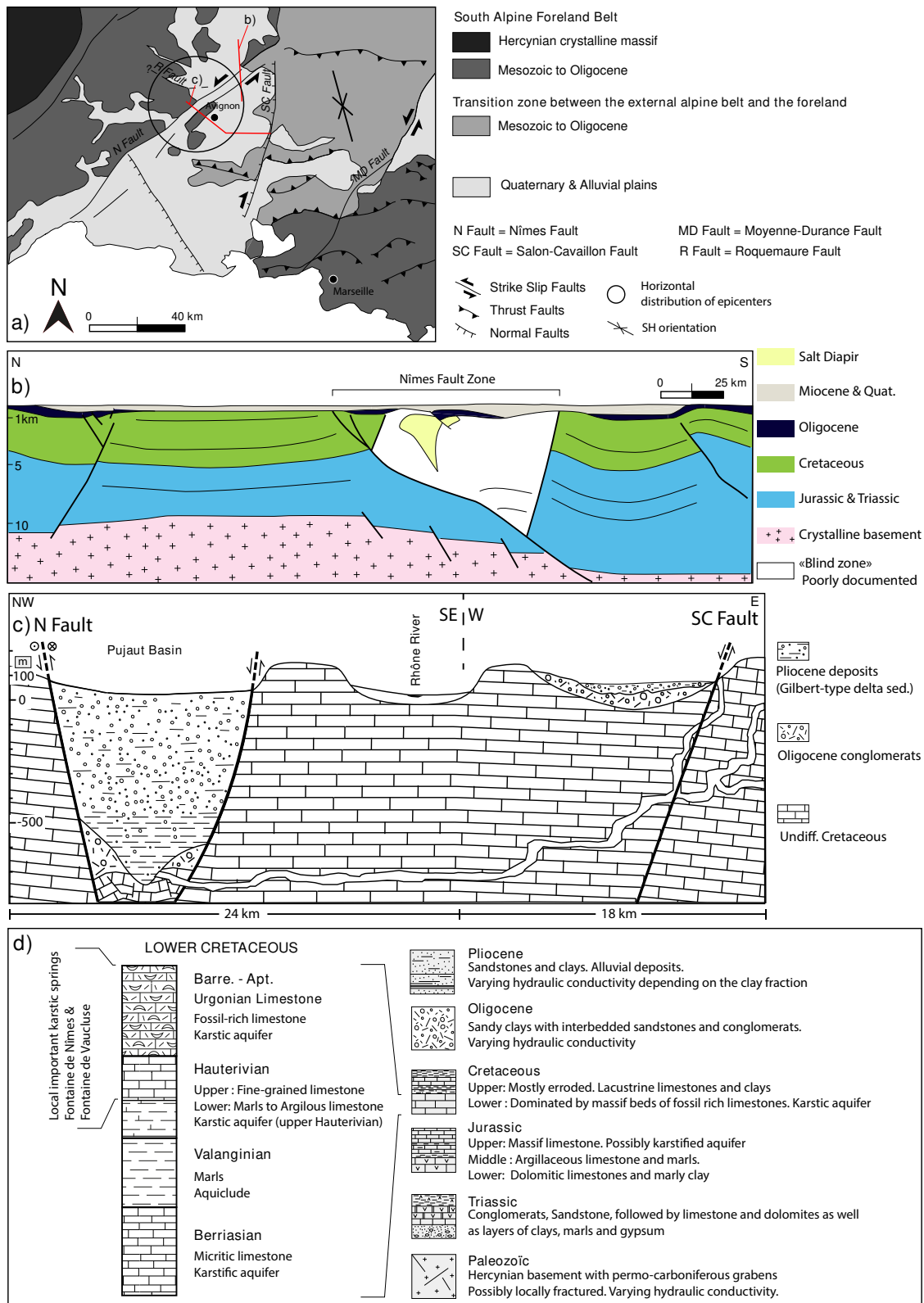


Figure 2.6: a) Tectonic map of the Provence area with the major deformation structures and location of b) and c). b) Simplified cross section over the Nîmes fault zone, c) schematic figure of the Pujaut Basin and the development of karst network during the Messinian crisis, and d) lithologies and their hydraulic properties.

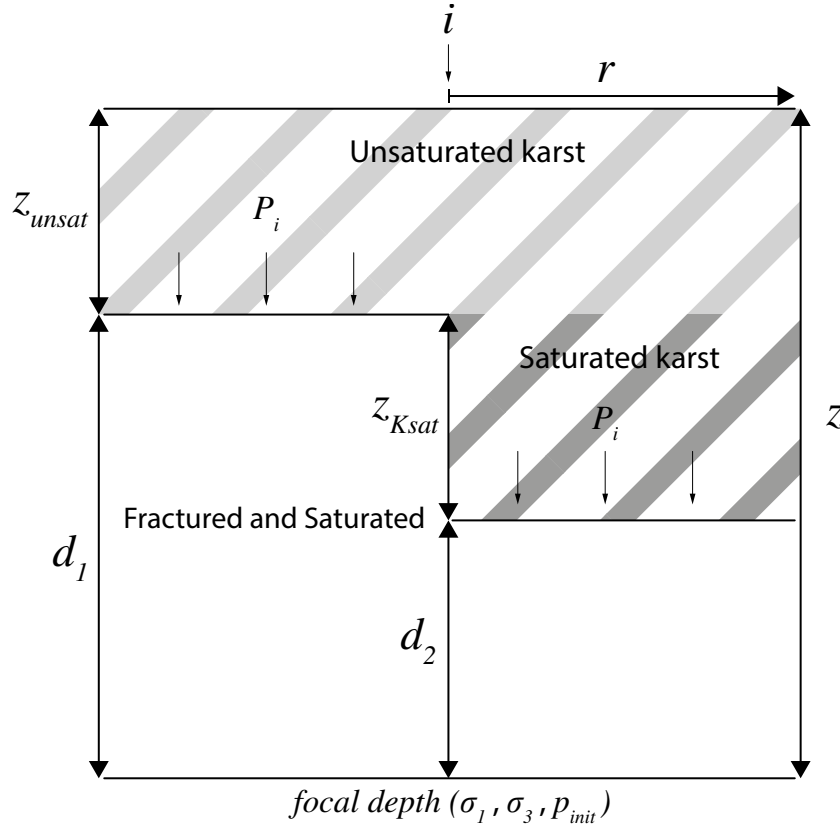


Figure 2.7: Considered environment for the analytical modeling of pressure front diffusion in a karstic area. A two layered semi-infinite environment of thickness z with initial stress conditions of σ_1, σ_3 and the initial pore pressure p_{init} at depth z . On top of this semi-infinite environment, a circular area of radius r is loaded with water from the precipitation of intensity i . P_i corresponds to the pressure induced from the load. d_1 and d_2 correspond to the distance the pore pressure front induced by P_i has to migrate to reach focal depth and depends on the presence or not of a saturated karstic network below the permanent water table. Arrows indicate the layer interface at which the onset of pore pressure diffusion takes place. See text for further explanations.

karstic channels. For the delayed effect, simplified drained conditions are considered and represented by a linear decrease occurring over 12 days from the maximum pore pressure increase induced by Δh at time $t = 0$ to 0 at time $t = 12$ days. The diffusion front is considered to migrate from the bottom of the karstic network towards focal depths. The distance d varies according to the extent of the karstic network below the water table. Two different values are tested, the first one, d_1 , considers that the pressure front starts migrating from the top of the saturated thickness, meaning that the distance d_1 corresponds to the total depth z minus the unsaturated thickness z_{unsat} , the second, d_2 considers a saturated karstic network with a thickness z_{Ksat} of 1000 m, meaning that the distance d_2 corresponds to the total depth z minus the unsaturated thickness z_{unsat} minus z_{Ksat} (Figure 2.7). Note that the analytical modelling aims at comparing the mechanisms to each other, not to do a sensitivity analysis of the different parameters.

Table 2.1: Input values for the analytical modelling. The initial stress conditions are calculated for the focal depth.

Case study	Hochstaufen	Muotathal	Avignon
Mean altitude at epicentre location [m.a.s.l.]	1500	1400	200
Average focal depth [m.b.s.l.]	1000	2000	3000
Total depth z [m]	2500	3400	3200
Base water level [m.a.s.l.]	1135	638	10
Unsaturated thickness z_{unsat} [m]	365	762	190
σ_v [MPa] at depth z	56.4	76.7	72.2
σ_H [MPa] at depth z	56.4	76.7	72.2
σ_h [MPa] at depth z	33.4	42	43.5
Initial pore pressure p_{init} [MPa]	20.1	25.8	29.4
Cohesion c [MPa]	0	0	0
Coefficient of friction μ [-]	0.6	0.6	0.6
Radius r [km]	5	2.5	19
Considered rainfall intensity $i_1; i_2$ [mm/d]	140; 130	100; 100	320; 80
Storage coefficient S [-]	0.001	0.001	0.001
Skempton coefficient B [-]	0.8	0.8	0.8
Poisson's ratio ν [-]	0.3	0.3	0.3
Diffusivity D [m ² /s]	1.0	1.0	1.0

2.5 Results and Discussion

Figure 2.8 shows the impact of a heavy rainfall on pore pressure through a direct hydrogeologic connection, poro-elastic deformation and diffusion. For the first two processes, the maximum increase occurs rapidly after the precipitations, reaching values of the order of MPa for the direct connection and only kPa for the poro-elastic deformation. Pore pressure increase resulting from the diffusion process varies over two orders of magnitude, depending on the distance travelled by the diffusion front. The maximum increase is delayed in regards with the precipitation event. A successive rainfall, even though of smaller intensity, results in an even higher pore pressure change due to the already accumulated water in the system. Thus, not only is the intensity of a given rain event important, but also the history of precipitation preceding the event.

Table 2.2 shows the maximum values of pore pressure increase as well as their corresponding CFS changes. Changes in CFS are positive for each process, but have different orders of magnitudes: from kPa for the vertical stress increase, the poro-elastic deformation, towards the order of MPa when considering the diffusion process, and clearly in the order of MPa for the direct connection.

Based on the timing of the beginning of seismicity at the three test sites with regards to the increasing pore pressure resulting from the different mechanisms, we suggest that seismicity is highly influenced by the direct effect of an increasing hydraulic gradient (direct connection) for all three cases, as seismicity occurs very shortly after the maximum rain intensity. However, the diffusion process can still be considered: setting a higher diffusivity and a smaller distance drastically impacts the response of pore pressure by increasing the amplitude of the change and reducing the time for the pressure front to reach focal depth. Thus, by adjusting the diffusivity and the distance, the response of the diffusion process could also correspond to the beginning of

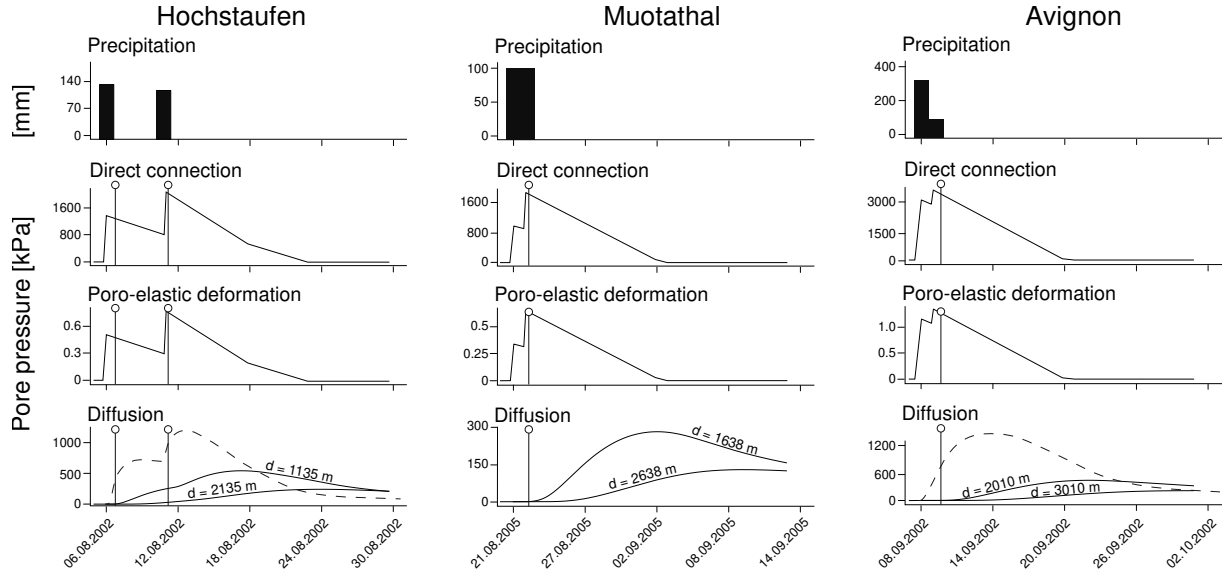


Figure 2.8: Considered rainfall intensity for the analytical modelling and the evolution of pore pressure changes over time for the different mechanisms, considering a linear decrease over 12 days for the system to drain the infiltrated water. Lines with an open circle at the top represent the onset of seismicity. Dashed line in first and third case study shows the response in pore pressure considering a hydraulic diffusivity of $10 \text{ m}^2/\text{s}$ and the shorter distance for both cases (1135 and 2010 m, respectively).

Table 2.2: Estimated initial conditions using the average lithostatic and hydrostatic pressure gradients, results of the direct and delayed effects as well as their impact on the CFS.

Case study	Hochstaufen	Muotathal	Avignon
Initial conditions			
Effective maximum stress σ'_1 [MPa]	35.5	50.9	42.7
Effective minimum stress σ'_3 [MPa]	12.5	16.2	14.0
Effective shear stress τ' [MPa]	9.9	15.0	12.4
Initial CFS [MPa]	-1.0	-0.1	-0.3
Considered rainfall intensity $i_1; i_2$ [mm/d]	140; 130	100; 100	320; 80
Direct effects			
Water head increase ΔH [m]	140; 210	100; 192	320; 373
Direct connection dp [MPa]	1.37; 2.06	0.98; 1.87	3.13; 3.65
Direct connection ΔCFS [MPa]	+0.82; +1.24	+0.59; +1.12	+1.88; +2.19
Vertical stress increase $\Delta\sigma_v$ [kPa]	1.37; 2.07	0.91; 1.74	3.14; 3.66
Poro-elastic deformation dp [kPa]	0.51; 0.77	0.34; 0.65	1.17; 1.36
Poro-elastic deformation ΔCFS [kPa]	+0.30; +0.46	+0.20; +0.38	+0.7; +0.82
Delayed effect			
Diffusion dp			
$d_1 = z_{\text{sat}}$ [m]	2135	2638	3010
Maximum dp [MPa], days [d]	0.24, 18	0.13, 19	0.20, 23
ΔCFS [MPa]	+0.15	+0.08	+0.12
$d_2 = z_{\text{sat}} - 1000$ [m]	1135	1638	2010
Maximum dp [MPa], days [d]	0.55, 11.5	0.28, 12	0.42, 14
ΔCFS [MPa]	+0.33	+0.17	+0.25

seismicity (dashed line in Figure 2.8). We propose that for the case of Muotathal and Avignon, the direct connection dominates the triggering mechanisms, as rupture occurs only hours after the maximum pore pressure is reached (1.8 and 3.65 MPa, respectively). For Hochstaufen, we suggest a coupled mechanism between the direct connection and the diffusion process, where the pore pressure at the bottom of the karstic network induced by the direct connection (up to 2 MPa) diffuses towards focal depth and triggers seismicity approximately one day after the rainfall, when the pore pressure increases of 0.5 and 0.8 MPa, for the first and second rainfall, respectively. For the three investigated case studies, authors have determined time lags between the rainfall event and the seismicity. These time lags are often used for comparison when discussing meteorological triggered seismicity. As Figure 2.9 shows, these time lags do not correspond to the same time period in each case. For Germany the time considered is between the onset of the first important rainfall and the maximum increase in seismicity 10 days later. For Switzerland, it is the time between the onset of rain and the beginning of seismicity and for France it is the time after the end of the storm and the beginning of seismicity. The case of Germany (Kraft et al., 2006b), however, shows that seismicity increases directly after the first important rainfall on the 6th of August. After the second rainfall, on the 11th of August, seismicity increases again to reach its maximum around the 16th of August. For Switzerland, the beginning of the rainfall is considered to be 4 days before the beginning of seismicity, however, Figure 2.9 clearly shows that more than 80% of the total precipitations fell during the two last days. Considering a same time period for each case study, for example the time between the onset of a heavy rainfall and the beginning of seismicity, leads to time lags that are comparable, between 24 and 48 hours. These time lags are well below the times observed for the delayed effects (Table 2.2), but as already mentioned, a change in diffusivity and distance can lead to comparable time lags. Considering the direct effects, one would expect seismicity to occur simultaneously as the peak rain intensity, however a time lag could result from the time needed for the water to be drained towards the water table. A combination of both, the effect of a direct hydraulic connection towards important depth followed by the diffusion process over a relatively small distance could also explain a smaller time lag.

However, setting time lags between the beginning of rainfall and seismicity is not optimal, since each region has its own drainage capacity, depending on the amount of soil and the epikarst thickness, the maximum head increase will not be reached at the same time after a rainfall. To better evaluate the impact of each mechanism on fault's stability for each case study and to simplify comparison, considered time lag should be taken between the maximum water head increase and the beginning of seismicity. Yet, such data are often difficult to collect, but the time at which a nearby karstic spring reaches its maximum discharge could be taken as a proxy, as the discharge is directly related to the hydraulic head. The case of Muotathal offers good data to estimate the fluid pressure when seismicity started using spring discharge values and hydraulic head variation observations (Figure 2.10). The region presents a karstic spring at 635 m.a.s.l. draining the Hoelloch and Silberer karstic system (Jeannin, 2001). During the heavy rainfall of 2005, a cave located 540 m above the permanent water table was flooded (Miller, 2008), implying an increase of about 5.4 MPa of fluid pressure. The normal water table fluctuation due to seasonality is between 40 m near the spring to approximately 250 m at 5'200 m distance

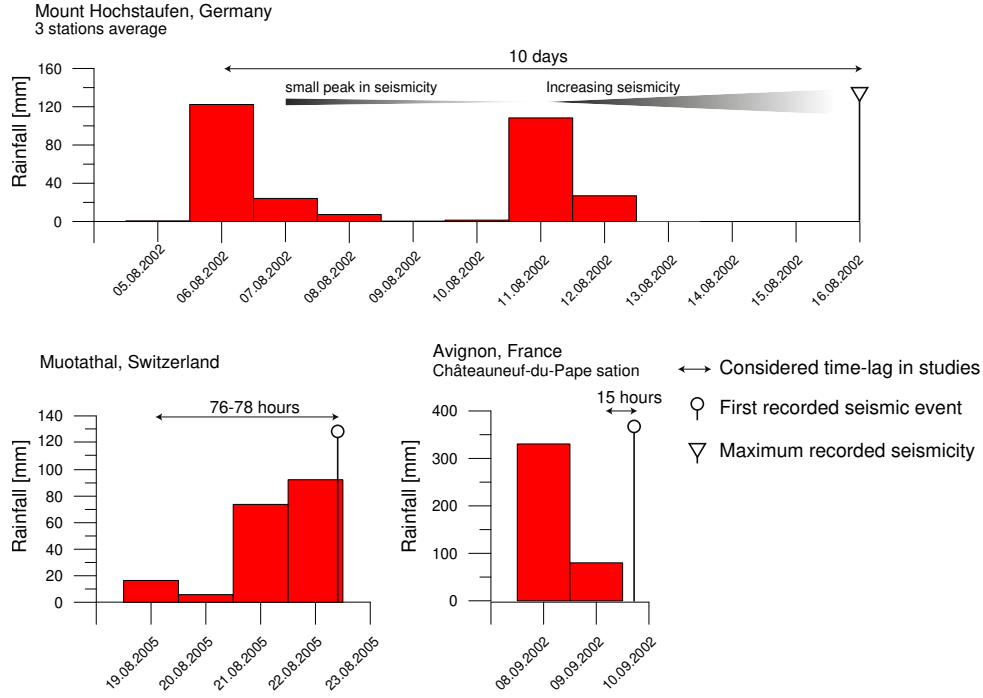


Figure 2.9: Abnormal rainfall event for Germany, Switzerland and France. Also indicated the maximum recorded seismicity for Hochstaufen (Germany), the first recorded events for Muotathal (Switzerland) and Avignon (France), as well as the time lag considered by the authors of the case studies.

from the spring (Jeannin, 2001), meaning a maximum variation of about 2.5 MPa between low waters and high waters during a hydrological year. Consequently, 5.4 MPa corresponds to an increasing pressure of 2.9 MPa above the maximum *usual* high water pressure.

The time at which the cave located at 540 m above the permanent spring was flooded is unknown and cannot be associated with a discharge value at the Schlichenden Brunnen. However, assuming that this cave was flooded during the maximum discharge rate at $22 \text{ m}^3/\text{s}$ (Figure 2.10), and that during the average high waters the level is around 250 m at 5'200 m from the spring, with a discharge of $12 \text{ m}^3/\text{s}$ (Swiss Federal Office for the Environment, 2017), it means that the critical value of the hydraulic head, i.e. the level at which shear slip is enhanced in the faulted system, is between 250 m and 540 m above the low water table.

Considering the data from Muotathal (Husen et al., 2007) seismicity starts when the discharge value reaches $16 \text{ m}^3/\text{s}$ (Figure 2.10). In order to have a rough estimation of the water column at this discharge rate, an experimental curve using three levels has been calculated. Using the values of 540 m, 250 m and 25 m with respectively the discharge values of 22, 12 and $3 \text{ m}^3/\text{s}$ (25 m and $3 \text{ m}^3/\text{s}$ values are taken from Jeannin (2001)) the following experimental function is found :

$$h = 4.6162Q^{1.5589} \quad (2.14)$$

For a discharge value of $16 \text{ m}^3/\text{s}$, and according to the relation shown in Eq. (2.14), the water level would be around 345 m above the *usual* low water level and consequently 95 m above the high water level (250 m) at 5'200 m from the spring. In terms of increasing fluid pressure it

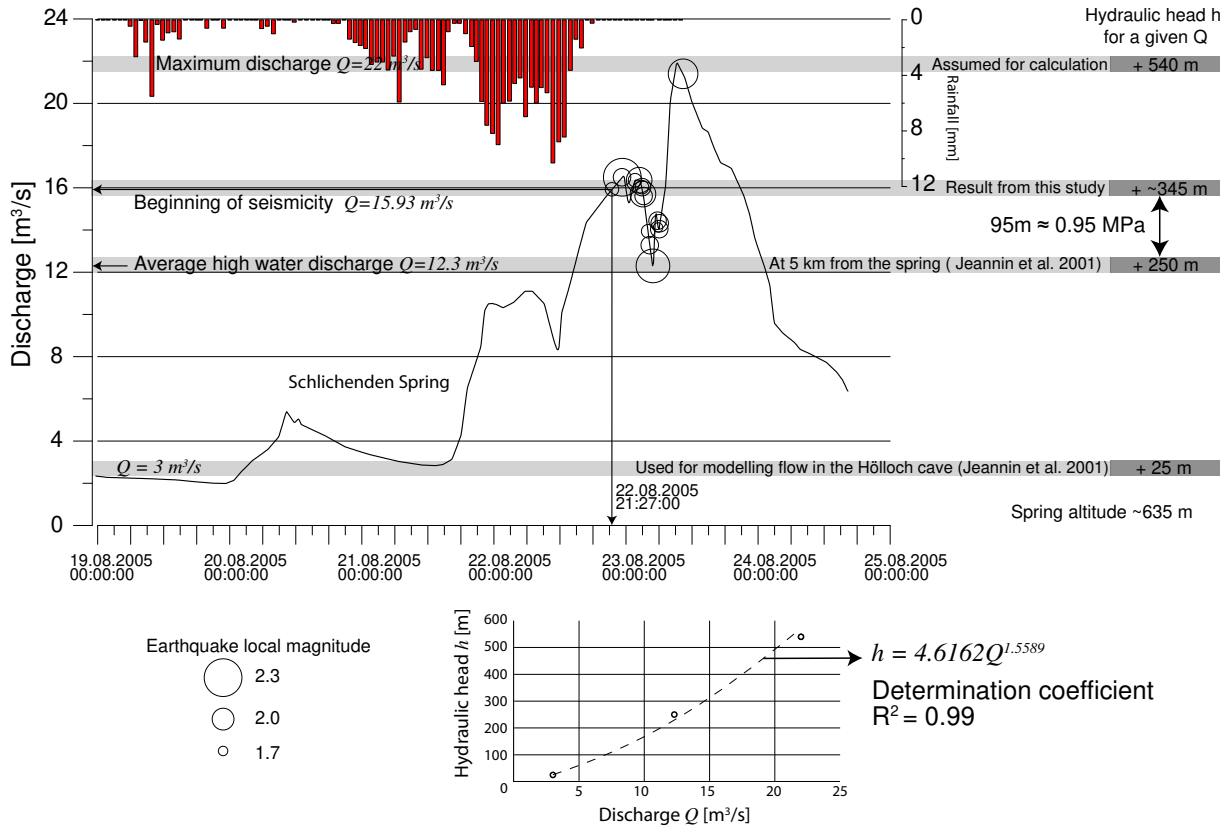


Figure 2.10: Schlichenden spring hydrograph, precipitations (red bars) and observed seismicity at specific discharge values (circles on the hydrograph). Highlighted in grey, the flow rates corresponding to specific water levels, indicated on the right side of the figure. The water levels +25 m, +250 m and +540 m are used for the elaboration of the rating curve (bottom) to get a relation between the flow rate Q and the water level h . The equation of the curve is used to calculate the water level at a given discharge. Also shown, the determination coefficient R^2 .

represents an increase of 0.95 MPa. This result assumes that during the maximum discharge of $22 \text{ m}^3/\text{s}$, the groundwater level was at its maximum at +540 m above the altitude of the spring (635 m), which means that this result is a minimum. Important to note is that this value is an estimation based on a relatively poorly constrained rating curve (3 values). Supplementary discharge values corresponding to specific groundwater level is necessary to improve this estimation. However, according to this simple analysis, seismicity starts when the critical value of +0.95 MPa is reached, which is of the same order of magnitude as the results obtain from the analytical modelling, where seismicity occurs following a pore pressure increase of $\approx 1.8 \text{ MPa}$. Literature suggests that the necessary stress changes to trigger seismicity on pre-existing faults can be as low as 0.01 to 0.1 MPa (Lockner and Beeler, 1999; Roeloffs, 1988; King et al., 1994; Saar and Manga, 2003; Hainzl et al., 2006). The upper bound of these values (10-100 kPa) are found in studies of triggered seismicity resulting from wastewater injection (e.g. Keranen et al. (2014); Stokes et al. (2023)). Lower values (1-10 kPa), are similar to stress changes resulting from earth tides effects (e.g. Tolstoy et al. (2002); Scholz et al. (2019); Zaccagnino et al. (2022)). However, although a correlation has been demonstrated between earth tides and earthquakes globally (Métivier et al., 2009), the high frequency of earth tides is usually inconsistent with the occurrence of seismicity locally, as demonstrated by Figure 4 in Hainzl et al. (2013). In that

sense, we suggest that changes of the order of the kPa are insufficient to trigger seismicity, which implies that changes in vertical stress as well as poro-elastic deformation due to the increasing load of water can hardly be considered as the triggering process for rain-induced seismicity. Nonetheless, modelling of seismicity in karstic regions has been performed previously, especially for the case of Hochstaufen, Germany (Hainzl et al., 2006, 2013). Authors found that seismicity was very well correlated with heavy rainfalls, but values found to trigger seismicity were of the order of the kPa, which is 3 orders of magnitude lower than values found in the present study. The major difference between results of Hainzl et al. (2006) and our study, principally lies in the difference in magnitude of the input signal used to model pore pressure variation at depth. For the diffusion process, we suggest that the pressure step increase is equivalent to the increasing hydraulic head in the karst conduits (100 m of water level increase $\approx$ 1 MPa of pore pressure increase), and that the distance the front has to migrate is the distance from the bottom of the karstic network (which can be several hundreds of meters below the surface) towards focal depth. Hainzl et al. (2006) use the precipitation data as input signal (mm of water) and the distance the front has to migrate is between the surface and focal depth. The difference in magnitude of the input signal explains the major difference between our results and theirs. Additionally, considering a shorter distance in our study, also influences the magnitude of pore pressure change (greater changes at shorter distances). The study of Hainzl et al. (2013), considers a possible water accumulation. They observe that an amplification factor of the rainfall signal of at least 10 (simulating water accumulation in open fractures) improves their model. These results corroborate the idea that water accumulation within karstic channels must be considered when modelling seismicity in karstic regions. We suggest that the amplification factor of the rainfall signal should be at least 1000 to correspond to the observed hydraulic head increase in karst conduits following important rainfalls.

2.6 Conclusions

From the simplified analytical model tested on three different case studies, neither the increasing vertical load nor the pore pressure increase due to the poro-elastic deformation are considered to be sufficient to trigger seismicity following heavy rainfall. Considering the channelling effect of a karst network in direct connection with focal depth, the effect of several tens of meter increasing water table impacts rapidly the pore pressure in the range of MPa. A small time lag between the beginning of the rainfall and seismicity could be explained by the time needed for the system to drain water towards the water table. If the karstic network is not in connection with focal depth, the diffusion process takes over and changes in pore pressure are attenuated over time and distance, but still possibly reaching the range of MPa. Figure 2.11 illustrates the final conceptual model and the most important parameters to consider such as focal depth, water table elevation and the extend of the karstic network below the water table. To trigger seismicity after an intense rainfall event in a karst environment, we suggest that either the karst network is in direct hydrogeologic connection with focal depth (Figure 2.11a), or relatively close to (Figure 2.11b), such that the pressure front reaches rapidly focal depth due to pressure diffusion and that the amplitude in pore pressure change is sufficiently large. To accommodate for a relatively small time lag, we propose to consider the superficial layer as having a possible

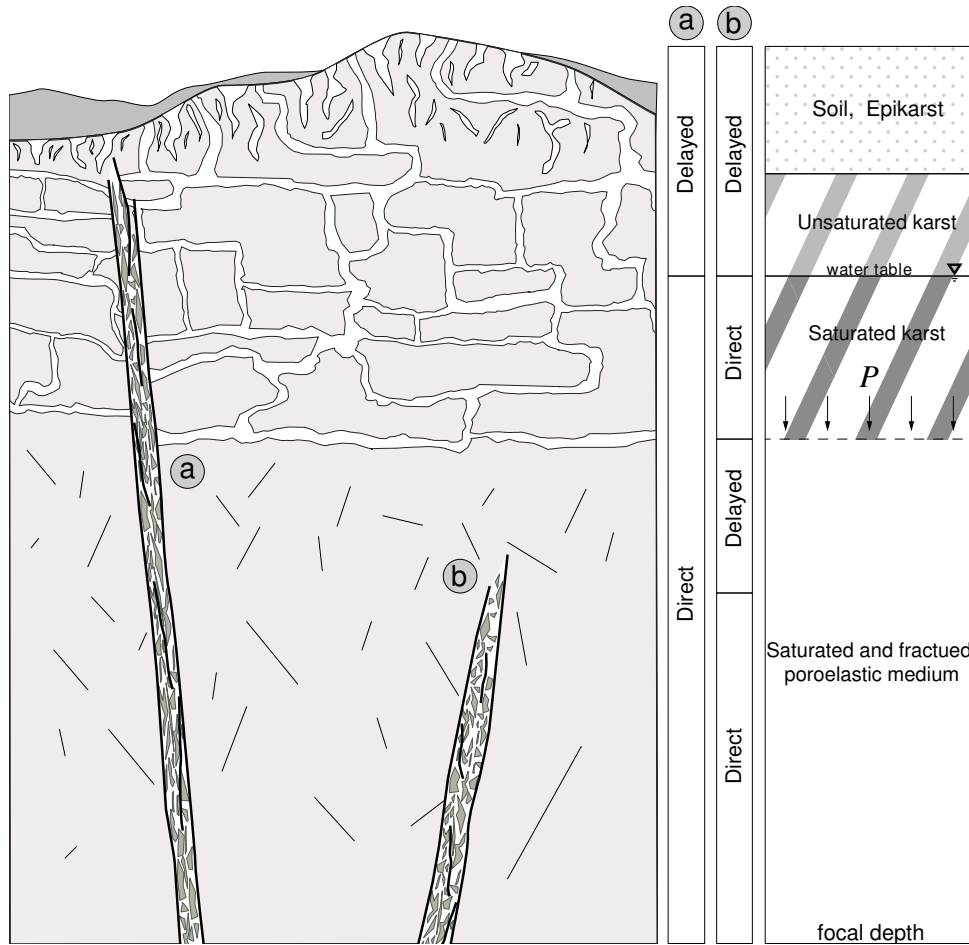


Figure 2.11: Final conceptual model considering the different layers and their impact on the pressure propagation towards focal depth, whether the pressure is rapidly (direct effect) or slowly (delayed effect) transmitted towards focal depth. a) Direct connection between the karst network and the fault zone. Drainage of the superficial layers (soil, epikarst and unsaturated karst) may act as a delaying factor until the water reaches the water table, after what, the effect is quasi instantaneous at focal depth if the fault acts like a conduit. b) No direct hydrogeologic connection between the fault zone and the karst network. The superficial layers may act as a delaying factor as in a). Once the water reaches the water table, the effect of the increasing water column is instantaneous at the bottom of the karst network. From there, the diffusion process takes over which acts as a delaying factor. Once the pore pressure front reaches the fault zone, which possibly acts like a conduit, pressure is transmitted rapidly towards focal depth.

delaying effect depending on its structure and composition. We suggest that the mechanisms behind seismicity following heavy rainfalls in karstic regions can be inferred from the observed time lag between the maximum discharge rate and the beginning of seismicity: no time lag to a time lag of a few hours supposes a direct connection between the karstic network and the focal depth, whereas a time lag from a few hours to a few days suggests a coupled solution of rapid fluid pressure diffusion towards the bottom of the karstic network followed by the diffusion process prograding towards focal depth. Based on our analysis we conclude that the mechanisms behind the seismicity of the three case studies are mainly driven by the direct connection, especially for the cases of Muotathal and Avignon where seismicity occurs only hours after the end of the precipitations, suggesting well developed and well connected karst networks for both sites.

For Mount Hochstaufen, the direct effect might be coupled with the diffusion process since the beginning of seismicity is delayed of a day after the maximum rain intensity. Although pore pressure diffusion process alone is not ruled out as mechanism, we advocate that the important hydraulic head variation in a karstic network as well as the extend of the network towards depth are the most important parameters to consider. Hence having a deeper comprehension of the geological and hydrogeological setting is essential.

Chapter 3

Joint seismic and hydrogeological data of the Neuchâtel area: a study on rain-induced fault activation

3.1 Introduction

Cases of seismicity induced by groundwater recharge have been reported by several authors (Wolf et al., 1997; Saar and Manga, 2003; Bragato, 2021; Kraft et al., 2006b; Husen et al., 2007; Rigo et al., 2008). Seismicity occurring unambiguously shortly after above average rainfalls have been observed in Germany (Kraft et al., 2006b), Switzerland (Husen et al., 2007) and France (Rigo et al., 2008), all three are located in well-known karstic regions. Following the concept presented in Chapter 2, the presence of well developed karstic networks is considered as an important prerequisites to have seismicity following heavy rainfalls, where the rapid increase in the hydraulic head within the karst conduits leads to important pressure variation at depth. If karst channels are hydraulically connected with the seismogenic depth, the variation in the hydraulic head instantaneously influences the stability of the fault through varying pore pressure. If no direct hydraulic connection exists, the pressure front is thought to migrate from the bottom of the karstic network towards focal depth, leading to a time lag between the maximum hydraulic head increase and the maximum pore pressure increase at focal depth. If a fault is critically stressed, these varying pore pressures might trigger seismicity. In the case of a direct hydraulic connection, seismicity would occur rapidly after the maximum hydraulic head, whereas if there is no hydraulic connection, a time lag would be observed between the maximum head increase and the beginning of seismicity, representing the time needed for the pore pressure front to migrate to focal depth. In order to evaluate if the Jura Mountains are also prone to seismicity following heavy precipitations, field investigations were carried out during the years 2019 to 2021, consisting in a seismic and hydrogeologic monitoring. In addition to evaluating the possible effect of rainfalls on fault stability, these investigations can inform on groundwater circulations at depth and possibly identify seismogenic zones sensitive to hydraulic head changes. The following work presents the results of two years of seismic and hydrogeologic monitoring and the combination of the two data sets.

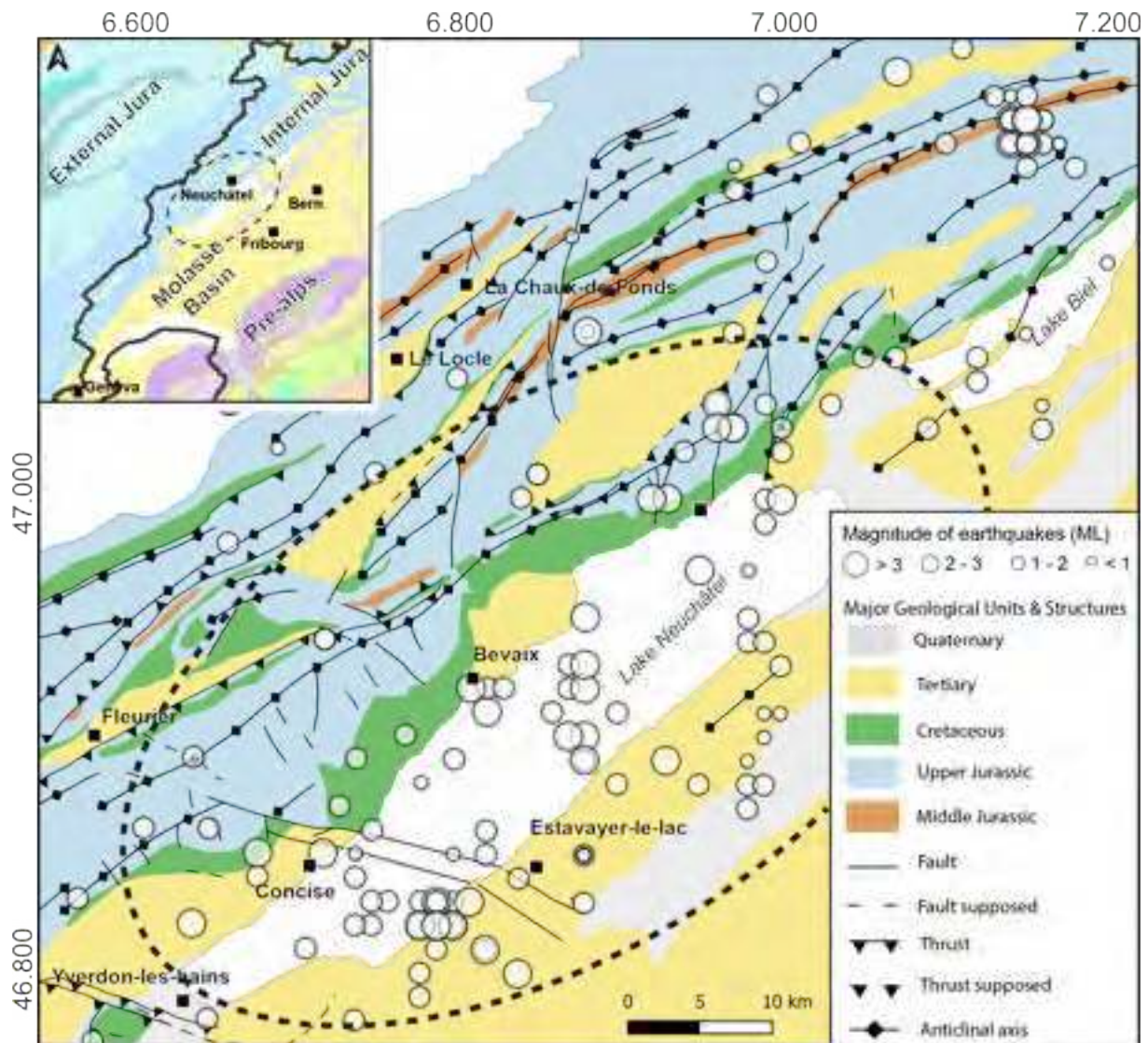


Figure 3.1: General situation of the study area (dashed line) and major tectonic units of western Switzerland, as well as major faults and thrusts (data from the Swiss Federal Office for Topography, swisstopo). White circles show the seismic events recorded by the Swiss Institute for Seismology (SED) between 2000 and August 2022.

3.2 Geological, hydrogeological and seismological context

The study area is located in the Internal Jura (Figure 3.1), on the northern shore of lake Neuchâtel (NW-Switzerland). The Jura Mountains are part of the Alpine orogeny and form a well-developed foreland fold-and-thrust belt that follows a thin-skinned deformation mode, in which the Cenozoic and Mesozoic sedimentary cover is detached from the underlying basement and pushed northward over a décollement level, identified as the Triassic evaporites (Burkhard and Sommaruga, 1998). The Jura is divided into the External Jura, composed of Plateaus separated from each other by narrow strongly deformed zones called the Faisceaux, and the Internal Jura, also called the folded Jura or "Haute Chaîne" (Sommaruga, 1997) (Figure 3.1). Folds of the Internal Jura are of relatively high amplitude and are related to NW and SW vergent thrusts of kilometric dip-slip displacement, resulting in a duplication of the entire Mesozoic

sequence. These thrusts cross-cut the entire sedimentary sequence, from the main décollement up to the surface (Sommaruga, 2011). ENE-WSW and NNE-SSW striking faults of local to regional importance cross-cut these SW-NE striking folds, showing normal, reverse or lateral offsets. Interpretation of seismic lines suggest no significant offsets of the base of the Mesozoic and the Pre-Mesozoic units from one side to the other of the faults, suggesting that these structures are restricted to the sedimentary cover (Sommaruga, 1997, 2011). Recent studies have shown that the deformation of the Jura is on-going and is related to thrusts seated below the décollement level suggesting a change in the deformation style, between thin-skinned to thick-skinned (Mosar, 1999; Ustaszewski and Schmid, 2007; Madritsch et al., 2010).

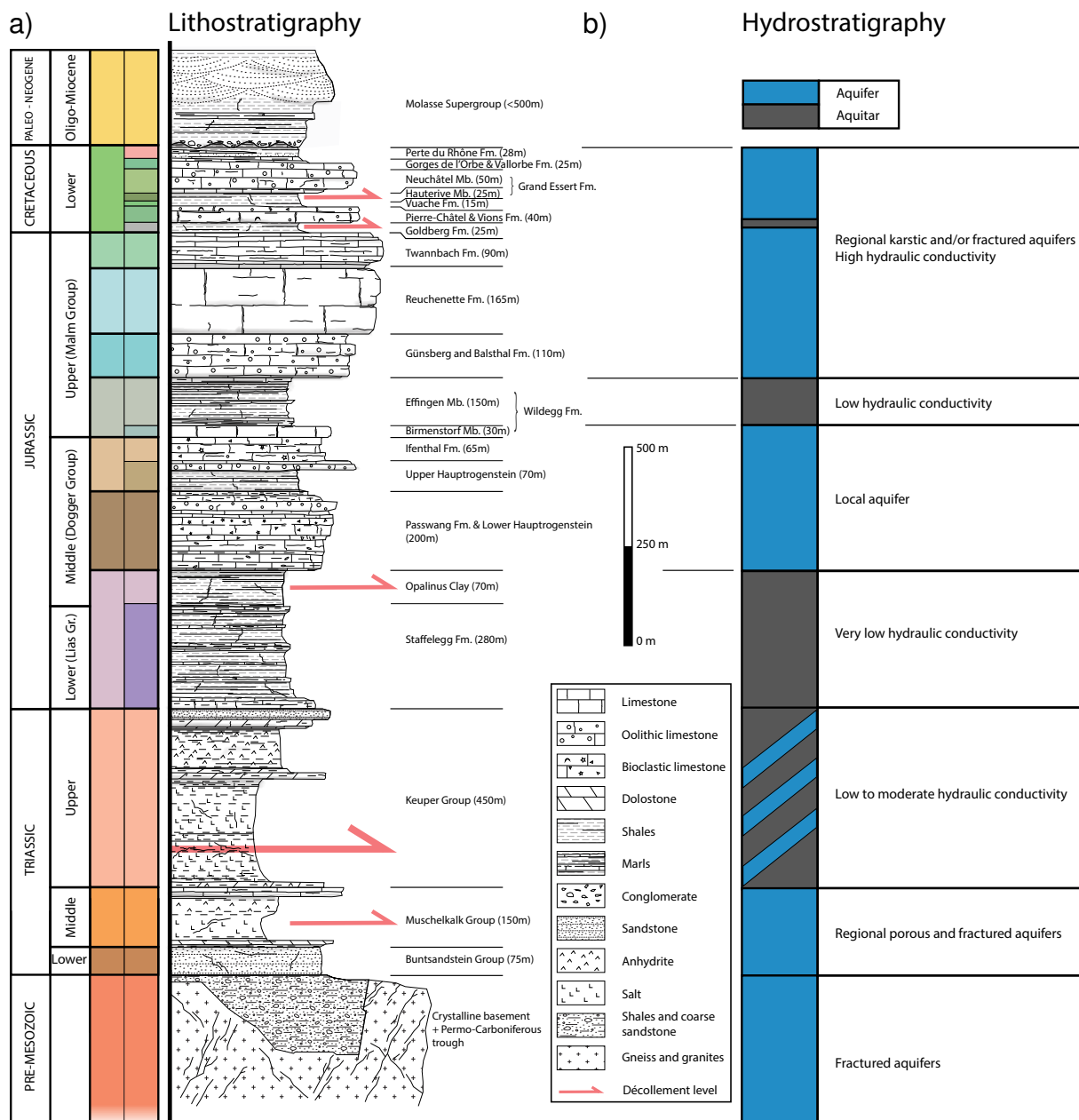


Figure 3.2: a) Lithostratigraphic log modified from Rime et al. (2019) illustrating the stratigraphic column of the Neuchâtel Jura and b) the hydrostratigraphic log illustrating the general hydrogeologic properties of the different lithologies.

The stratigraphic column of the study area presented in Figure 3.2a has been compiled by Rime et al. (2019) and illustrates the succession of lithologies. The undifferentiated Molasse series compose the upper part of the stratigraphic column and are deposited unconformably over the Lower Cretaceous. Composed of relatively thin layers of limestones alternating with more marly layers, the Cretaceous conformably deposited on the limestones of the Upper Jurassic, these formations constitute the major outcropping rocks in the study area. The bottom formation of the Upper Jurassic consists in the Wildegg Formation and is mostly constituted of marls with a total thickness of 180m. Underlying the Upper Jurassic (or Malm), The Middle Jurassic (or Dogger) consists in a succession of limestone beds with relatively small intercalation of marly layers. The bottom of the Dogger is composed of the Opalinus Clay, which is closer in lithology to the Lower Jurassic (or Lias), consisting mostly in clayey and silty deposits. Finally, the Triassic underlying the Jurassic and overlying the crystalline basement is mostly composed of evaporites such as halite and anhydrite in the Keuper and the Muschelkalk Groups (Upper and Middle Triassic) (Sommaruga, 2011). The Lower Triassic consists in layers of continental sandstones from the Buntsandstein Formation. The basement shows extensional deformation structure which are attributed to the latest stage of the Variscan orogeny, when crustal thinning took place in the Upper Carboniferous. These structures are called the Permo-Carboniferous trough or half-graben and are filled with detrital fluvial sediments (Sommaruga, 1997). The hydrogeology of the study area is dominated by the karst network that forms in the Cretaceous and Jurassic limestones. The Upper Jurassic (Malm) is the primary groundwater reservoir of the region. Other aquifers are found in the Cretaceous limestones and the Quaternary deposits. Springs associated with these reservoirs are of more or less importance, however, the major springs are outlets of the Malm. Although some springs are located in the Cretaceous, evidence through tracer testing (see e.g., Matthey (1976) and Appendix B) shows that major springs located in the Cretaceous are connected to the Malm reservoir, suggesting a connection between the two aquifers in some location (Király, 1973). The aquifers and the major impervious layers associated with the different lithologies are presented on Figure 3.2b. In comparison with other regions in Switzerland, the study area presents a low to moderate seismic activity. From 1600 to 1975, 4 earthquakes with magnitude 4.5 or above are thought to have occurred in the study area although location error are of several tens of km. From 1975, improvement of the seismic network allowed to detect more earthquakes of lower magnitude (Fäh et al., 2003). Between 1984 and 2001 a little over 30 events were recorded with magnitude of 2 or above (Ecos09 catalog, SED). From 2002 to August 2022 (Figure 3.1), a little under 140 earthquakes were recorded by the Swiss Seismological Service in the region of Lake Neuchâtel. Most of the earthquakes have magnitudes that range between 1.0 and 2.5. 1 event has a magnitude of 3.1 (near the village of Bevaix) and about 15 events show magnitudes lower than 1.0, down to 0.2. Although some major deformation structures are mapped in the region such as strike-slip and thrust faults, location of epicenters cannot be clearly linked to these features. Fault plane solutions are rarely identified due to the low magnitude of the events, making it difficult to link earthquake to specific deformation zones. Moreover, the error in the location or the presence of unknown subsurface structures hinders to confidently link events to known faults.

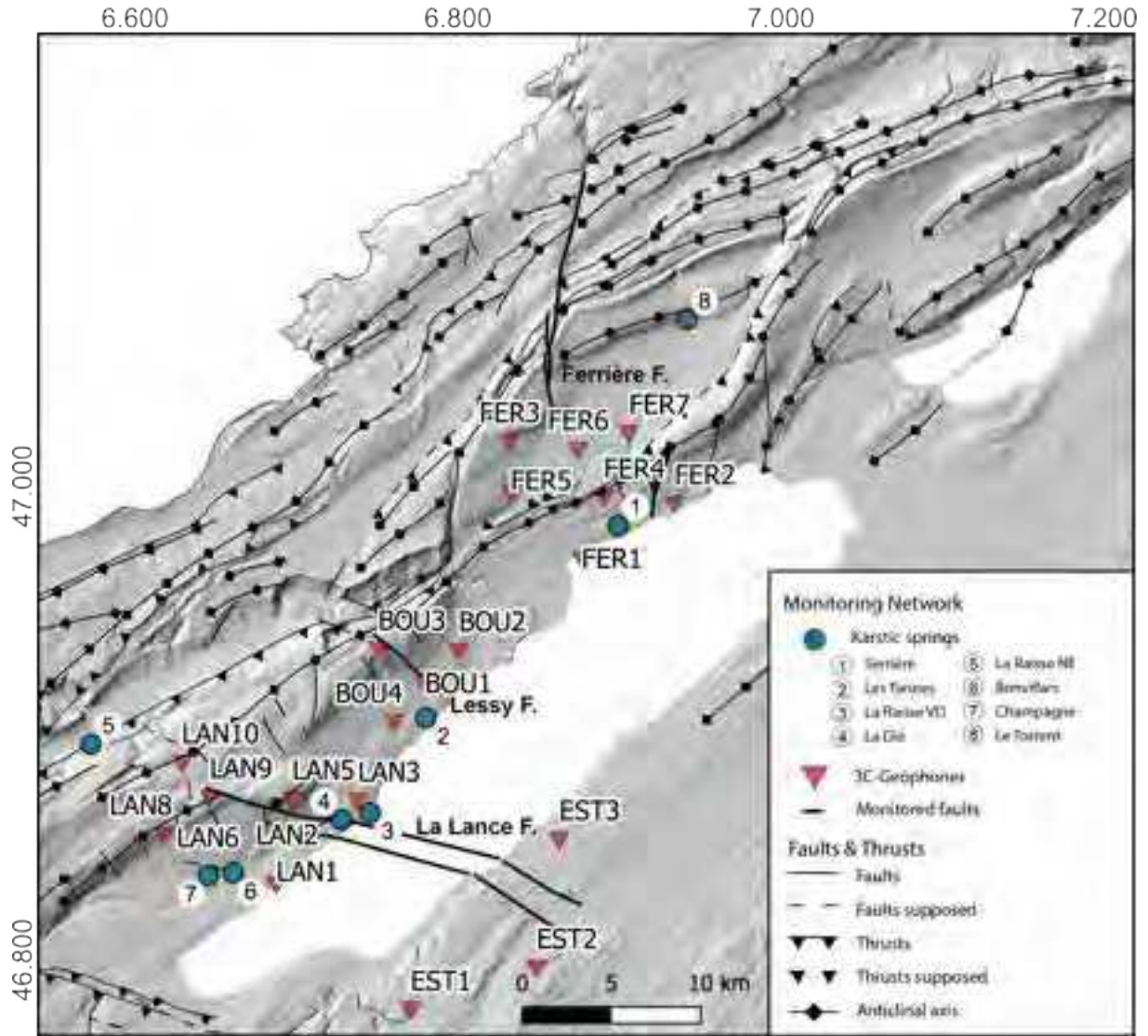


Figure 3.3: Seismic and hydrogeologic monitoring network around 3 major deformation structure, the La Lance, Lessy and Ferrière Faults. Digital elevation model (DEM) of the background map and major deformation structures are from the Swiss Federal Office for Topography, swisstopo.

3.3 Methods

Seismic monitoring network

For the acquisition of continuous seismic data, 3C ZLand nodes from Magseis Fairfield were installed around potentially active faults, the La Lance, the Lessy and the Ferrière Fault. Each node comprises 3 internal 5 Hz geophones in orthogonal position, a digitizer, a GPS, batteries and 32GB data storage. These instruments are cableless and can be buried under a thin layer of soil to reduce noise, as well as depredation. In order to have the best signal to noise ratio, the nodes must ideally be installed as far away from anthropogenic noise sources, such as main roads, railways, cities or quarries. Station coverage and accessibility must also be considered. These nodes run on batteries and data must be downloaded regularly: they must be replaced every 4 weeks, hence, compromises have to be found between the ease of access, the station

coverage and the amount of noise. The spatial distribution of the stations over the study area is shown on Figure 3.3. Stations are distributed over the northern and part of the southern shore of Lake Neuchâtel into 4 separate sub-networks, named according to their location (LAN and FER for the Lance and Ferrière faults, respectively, BOU and EST for the cities of Boudry and Estavayer-le-lac, respectively). The sampling interval for each node is set to 4 ms (250 Hz) in order to have enough storage capacity for the approximately 30 days life span of the batteries. In addition, data from the permanent network of the Swiss Seismological Service were also downloaded from EIDA (European Integrated Data Archive) infrastructure within ORFEUS (Observatories and Research Facilities for European Seismology).

Seismic data processing

The seismic data recorded on the 3C ZLand stations is downloaded using the working station that is provided with this monitoring system. This working station is called the Mini Seismic Management System on which different softwares are available to manage data collection and exportation. After data is downloaded, it can be exported under the Fairfield continuous format to be then converted into the more conventional miniseed file format. Day files containing data of all stations of one region (e.g. FER, BOU, LAN, EST) are created. The detection of earthquake is done using the STA/LTA trigger algorithm (see for example Trnkoczy (2009)). This method consists in scanning day files of one region with two time windows, a short time window (ST) and a long time window (LT). The amplitude of the data within these windows is averaged, resulting in a short time average (STA) and a long time average (LTA). The ratio of these two values, STA/LTA, is constantly compared to a threshold value set by the user. If this value is reached, it corresponds to a trigger, and data is extracted from the day files. Parameters used for the STA/LTA trigger algorithm are presented in Appendix D.1. Visual inspection of each trigger is necessary as false triggers are common. Once the data has been verified, the picking process can begin. Figure 3.4 schematizes the workflow from the node in the field to event identification, the phase picking and further seismological investigations.

While the conversion of the files and the detection of events is done using the ObsPy toolbox in python (Beyreuther et al., 2010), the validation of the triggers and the picking of the phases is done on the SeisAn software (Havskov and Ottemöller, 1999), a complete seismic analysis system. This software is used for phase picking, event location and magnitude calculation. When possible, phase picking is done on unfiltered data. A bandpass filter (1-30Hz) is applied if necessary. Picking is realized following the recommendations found in Diehl et al. (2012), and is done by a single operator to maintain consistency. A weight is attributed for each P and S pick and corresponds to the time error interval, which is the time between the earliest possible pick and the latest possible pick; the smaller the time error interval the higher the quality of the pick. Weighting classes range from 0 to 4, 0 having an error interval of 0.05 seconds and 4 an interval higher than 0.4 seconds. Picks attributed the weight of 4 are disregarded for further analysis. Picking is done in a 1 second time window for each phase. For the location, SeisAn uses a modified version of HYPOCENTER (Lienert et al., 1986). The P-wave velocities model used for this study (Table 3.1) is derived from the 3D velocity model of Switzerland (Husen et al., 2003), the S-wave velocities are calculated using a V_p/V_s ratio of 1.71.

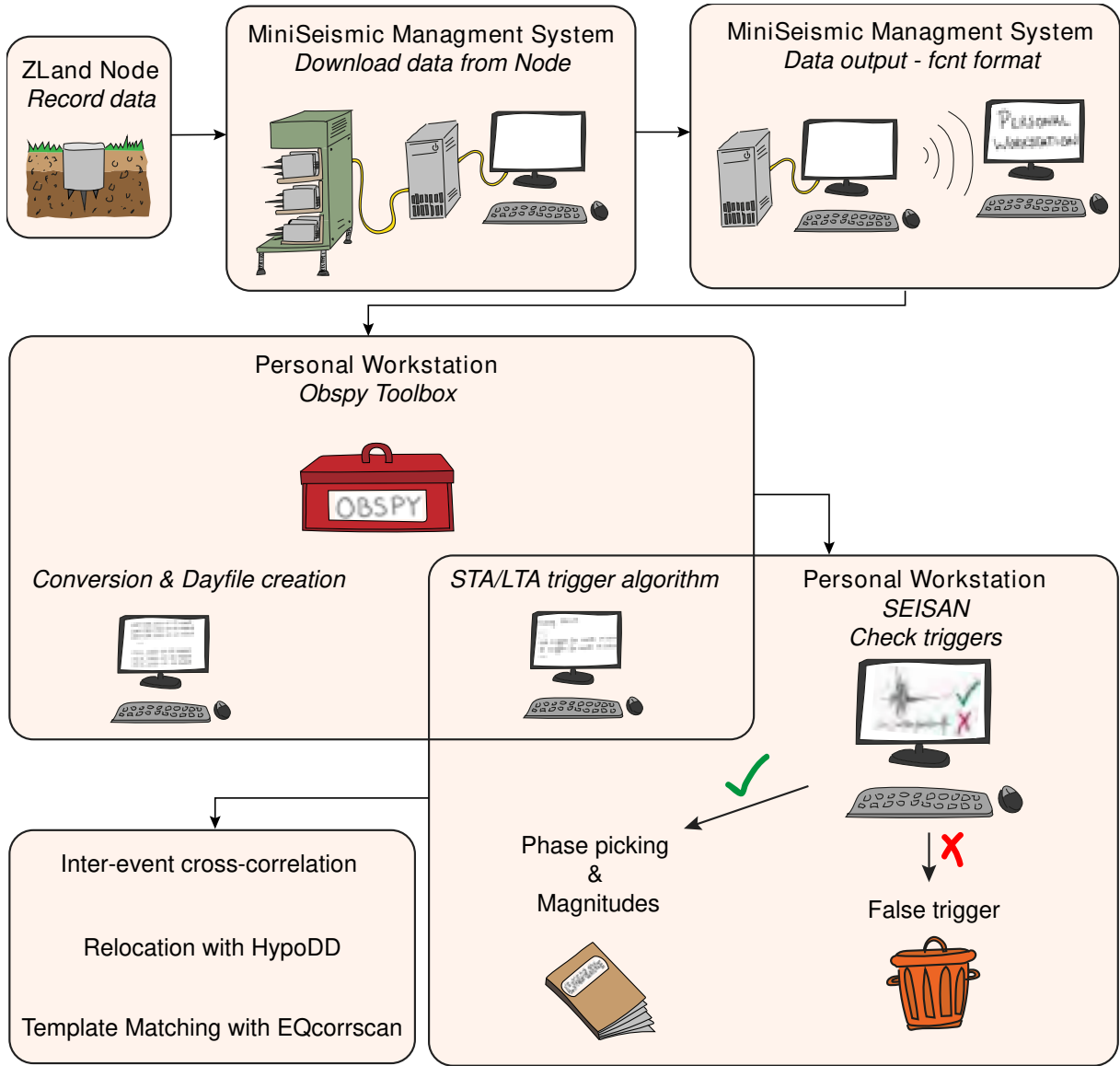


Figure 3.4: Schematic workflow for seismic data acquisition and analysis

To calculate magnitudes, a response file (RESP file) for each channel and each station is created using the PDCC (Portable data collection center) program. A RESP file contains both metadata and response characteristics to ground motion of an instrument. The RESP file removes the instrument response and is needed for local magnitude calculation since this type of calculation uses a Wood-Anderson instrument as reference. Local magnitudes have been calculated using the Swiss Seismological Service (SED) scale presented in [Edwards et al. \(2015\)](#) :

$$M_L = \log(amp) + C_d \quad (3.1)$$

where amp is the maximum mean to peak amplitude in mm read after applying a Wood-Anderson filter on the data and C_d is a distance correction and varies according to the distance between the station and the hypocentral distance. The distance correction C_d is given by [Edwards et al.](#),

2015):

$$C_d = 0.0180 \cdot R + 1.77 + 0.968 \cdot \log_{10}(R/20) \quad \text{for } R \leq 20\text{km} \quad (3.2)$$

$$C_d = 0.0180 \cdot R + 1.77 \quad \text{for } R \leq 60\text{km} \quad (3.3)$$

$$C_d = 0.0038 \cdot R + 2.62 \quad \text{for } R > 60\text{km} \quad (3.4)$$

where R is the hypocentral distance. Maximum amplitudes were read in SeisAn and magnitudes were calculated using Eqs. (3.1)-(3.4) implemented in the code of Antunes et al. (2020) to calculate magnitudes from a SeisAn catalog file.

Table 3.1: P-Wave velocities for the study area derived from the 3D velocity model of Husen et al. (2003). S-wave velocities are calculated using a V_p/V_s ratio of 1.71

Depth [km]	P-wave [km/s]	S-wave [km/s]
-5.0	4.60	2.69
0.0	4.60	2.69
2.0	5.98	3.49
5.0	6.00	3.50
8.0	6.03	3.53
15.0	6.09	3.56
22.0	6.16	3.60
30.0	6.24	3.65
40.0	6.33	3.70
50.0	8.10	4.74
60.0	8.15	4.77

Hydrogeologic monitoring network

Observation of the hydraulic head increase in the karstic network is not a trivial task and requires speleological knowledge. In order to evaluate the rising hydraulic gradient in a karstic network without necessitating specific skills, the flow rate was continuously monitored at major karstic springs. This approach does not give an absolute value but informs on the variation of the hydraulic gradient since the flow rate is directly related to it. The higher the flow rate, the higher the hydraulic gradient. In total, 7 springs were monitored (Table 3.2). Among them, 5 were monitored over approximately 2 years and 2 were monitored only the last 6 months of the observation period. The sampling rate of the level loggers installed at the spring was set between 1 sample per 15 minutes to 1 sample per hour depending on the spring. A supplementary observation point was made available by the Spéléo Club des Montagnes Neuchâteloise, who kindly shared their data of the water level within a sinkhole (point 8 in Figure 3.3).

Hydrogeologic data processing

The automatic level loggers installed at the different karstic springs measure the total pressure above the logger. This pressure comprises the water column as well as the atmospheric pressure. To get the pressure of the water column, the atmospheric pressure is subtracted to the total pressure. Atmospheric pressure is recorded by the Federal Office of Meteorology and Clima-

Table 3.2: Monitored springs, their coordinates, altitude, whether they are permanent or temporary and the parameters recorded : Δh is the water level fluctuation, T is the temperature and EC is the electrical conductivity

Springs	Lat.	Long.	alt.[m]	Regime	Parameters
La Serrière	46.9856	6.9013	466.8	Permanent	Δh , T , EC
Les Tannes	46.9048	6.7835	472.9	Permanent	Δh , T
- La Raisse VD	46.8631	6.7506	431	Permanent	Δh , T , EC
La Dia	46.8603	6.7322	469.8	Temporary	Δh , T
La Raisse Fleurier	46.8910	6.5776	757.4	Permanent	Δh , T , EC
Bonvillars	46.8377	6.6658	471.4	Permanent	Δh , T , EC
Champagne	46.8364	6.6506	483.1	Permanent	Δh , T , EC
Torrent	47.0726	6.9434	750.6	Temporary	Δh , T

tology MeteoSwiss at different location and can be accessed online for specific meteorological station (IDAWEB). For springs located too far away from these stations, pressure logger were installed near the springs. Once the atmospheric pressure is subtracted, the remaining pressure is converted into a water column. The variation of the water column directly reflects the variation in the discharge of the spring, the higher the water column, the higher the discharge. To convert the continuous water column measurements into discharge values, punctual flow rates measurements were realized during different hydrogeological periods associated to a specific water column. A best fitting rating curve is established between the flow rate and the water column and its function is used to calculate a discharge value for each value recorded by the level logger. Along with measuring the pressure, the level loggers also measure the water temperature, and, in some location, the electrical conductivity.

3.4 Results

Spring hydrographs

Among the 7 springs monitored over the period of observation (2019-2021), 2 permanent major springs are described hereafter, the Raisse VD and the Serrière. 2 temporary springs, the Dia and the Torrent, are also considered in this analysis, both of these springs are identified as overflows of the Raisse VD and the Serrière, respectively. Data of the Dia spring is from this study, whereas data of the Torrent has been collected and kindly shared by the Spéléo Club des Montagnes Neuchâteloises. Hydrographs of the other monitored springs can be found in Appendix C. Precipitations and snow cover information are from the Federal Office of Meteorology and Climatology (MeteoSwiss). Table 3.3 summarizes the key values mentioned in the text for the permanent springs.

Raisse VD

Figure 3.5 illustrates data from the Raisse VD springs from July 2019 to July 2021. Over the year 2019-2020, the discharge varies from a minimum of 130 l/s during the fall 2019 to a maximum of 6200 l/s at the end of the winter in 2020. Discharge values in between these minimums and maximums are punctually around 2500-3000 l/s following rainfalls and between

500 and 1500 l/s during recessions. In average the flow rate of the Raisse VD spring over the year 2019-2020 is around 1200 l/s. The water temperature varies approximately between 8.5 and 10.0° C. For the year 2020-2021, the minimum discharge seems to be between July and October 2020 around 670 l/s, however, the water temperature curve along with the sudden drop in electrical conductivity suggest that the logger was outside the water during this time period. Thus, the minimum discharge of the Raisse VD reached at least 670 l/s but could be lower. The maximum flow rate is around 7000 l/s in February 2021. Over the year, punctual maximums reach discharge values between 2000 and 4000 l/s after rainfalls and are between 1000 and 3000 l/s following recessions. In average the flow rate of the year 2020-2021 is around 1950 l/s. For the Raisse VD spring, two distinct dry periods are identified, both occurring during the summer, from July to October 2019 and 2020, with an average flow rate around 260 l/s for 2019 and 880 l/s for 2020. Between October and July 2019-2020 and 2020-2021, the hydrographs show several peaks and recessions, but the average flow rate during this period is noticeably higher than during the dry periods, around 1600 l/s for 2019-2020 and 2300 l/s for 2020-2021. In general, considering both dry and wet periods, the year 2020-2021 has higher discharge values. Periods of higher flow rates are directly related to precipitation events or snow melt periods. The time lag between the maximum intensity of rain and the maximum flow rate at the spring varies between 7 to 15 hours for floods occurring during the time period the spring was monitored. The decrease in the flow rate following the peak discharge is at first very rapid and then slows down over several days if no other relevant precipitation occur.

Serrière

Figure 3.6 illustrates data from the Serrière from July 2019 to June 2021. Over the period 2019-2020, the minimum measured flow rate is around 600 l/s at the end of the summer 2019 and in the spring 2020. The maximum flow rate is measured in December 2019 and reaches 12000 l/s. In between these minimums and maximums, the flow rate reaches punctual peaks between 6000 and 10000 l/s after rainfalls, followed by a rapid recession, at the end of which flow rates are close to the minimum. The water temperature lies between 7.4 and 8.7°C. For the year 2020-2021, approximately 3 months of data are missing between July and October 2020. The minimum measured flow rate for the time period where data is available is around 600 l/s and is reached several times over the year. The maximum flow rate reaches 14000 l/s in February 2021. Punctual discharge peaks reach values between 2000 and 7000 l/s followed by rapid recessions at the end of which the flow rate is close to the minimum. The water temperature lies between 7.9 and 8.9°C. During the year 2019-2020, two periods of several months show low discharge

Table 3.3: Minimum, maximum and annual average flow rates of the permanent springs of the Raisse VD and the Serrière. Also shown, annual average water temperature.

	2019-2020				2020-2021			
	Flowrate [l/s]			Temp. [°C]	Flowrate [l/s]			Temp. [°C]
	min.	max.	average	average	min.	max.	average	average
Raisse VD	130	6'200	1'200	8.98	670	7'000	1'950	9.18
Serrière	600	12'000	1'460	8	600	14'000	1'860	8.23

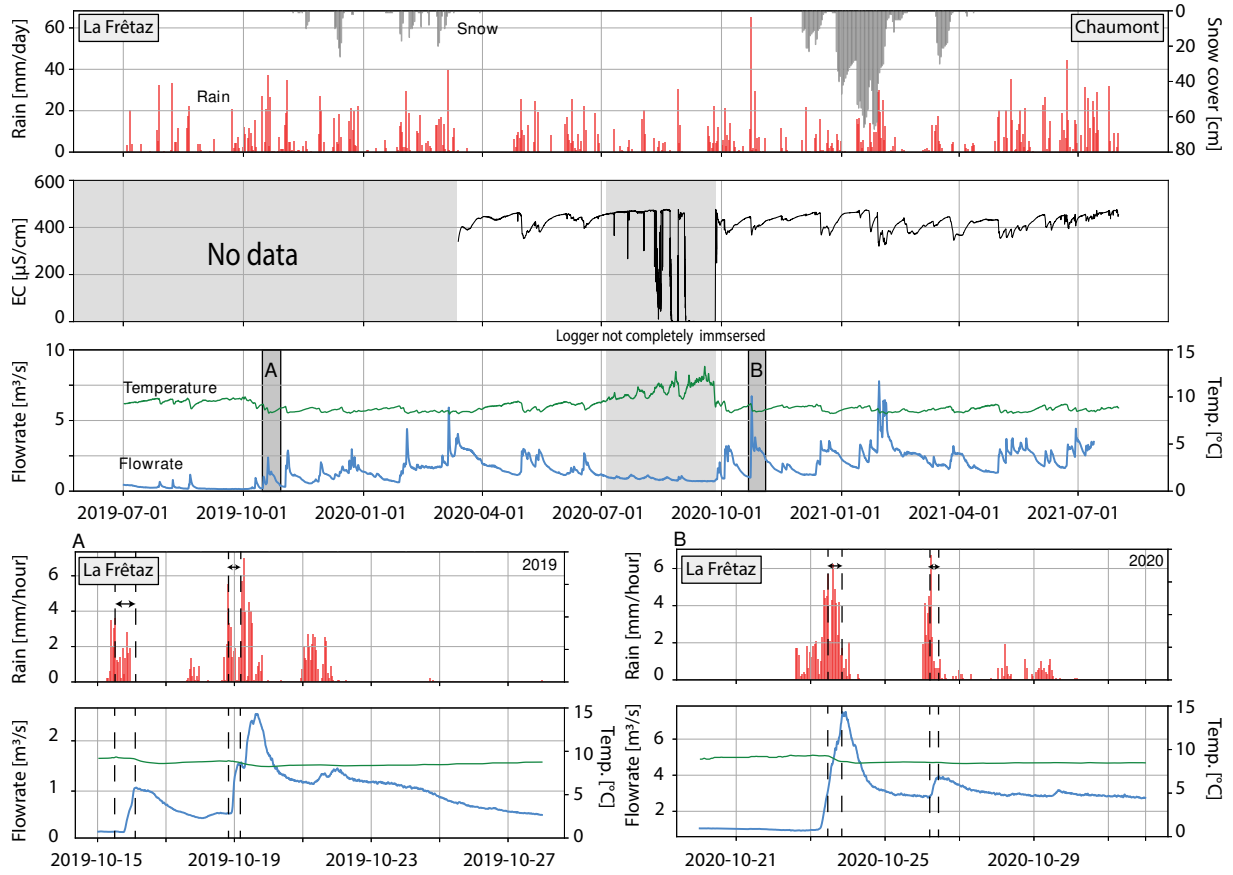


Figure 3.5: Hydrograph of the Raisse VD spring (blue line) from July 2019 to July 2021. Also shown, the rain (La Frêtaz station), the snow cover (Chaumont station) (red and grey bars, respectively), the temperature (green line) and the electrical conductivity (EC, black line) monitored since march 2020. From August to October 2020 logger was outside the water as attested by sudden drop in the electrical conductivity and the high amplitudes in water temperatures. Panels **A** and **B** show two distinct flood events in October 2019 and 2020, respectively. Dashed lines show the approximate time lag between the maximum rain intensity and its corresponding peak in the discharge curve.

values, from July to October 2019 and from April to July 2020, with average flow rates of 1000 l/s and 750 l/s, respectively. From October 2019 to April 2020 the frequency of higher flow rates increases. Over this time period, the average flow rate is 2000 l/s. For the year 2020-2021, two periods of rather low flow rates are identified, between November 2020 and January 2021, as well as between March and May 2021 with average flow rates of 900 l/s and 1000 l/s, respectively. November 2020, February 2021 and the period between April and June 2021, show higher flow rates with average values of 2900 l/s, 5100 l/s and 1500 l/s, respectively. Peaks in the discharge correspond to periods of snow melt or rainfall events. However, every precipitation period does not seem to have an impact on the flow rate, and the hydrograph's shape varies from one precipitation event to another, without consistency regarding the amount nor the duration of the rain. This is most certainly due to the hydroelectric power plant installed at the spring. Indeed, the water level from which the flow rate is estimated is measured in a retention basin collecting water from the spring. Part of the water flows directly to lake Neuchâtel, while part of it is channeled towards turbines. It is the water level in the basin that dictates the opening

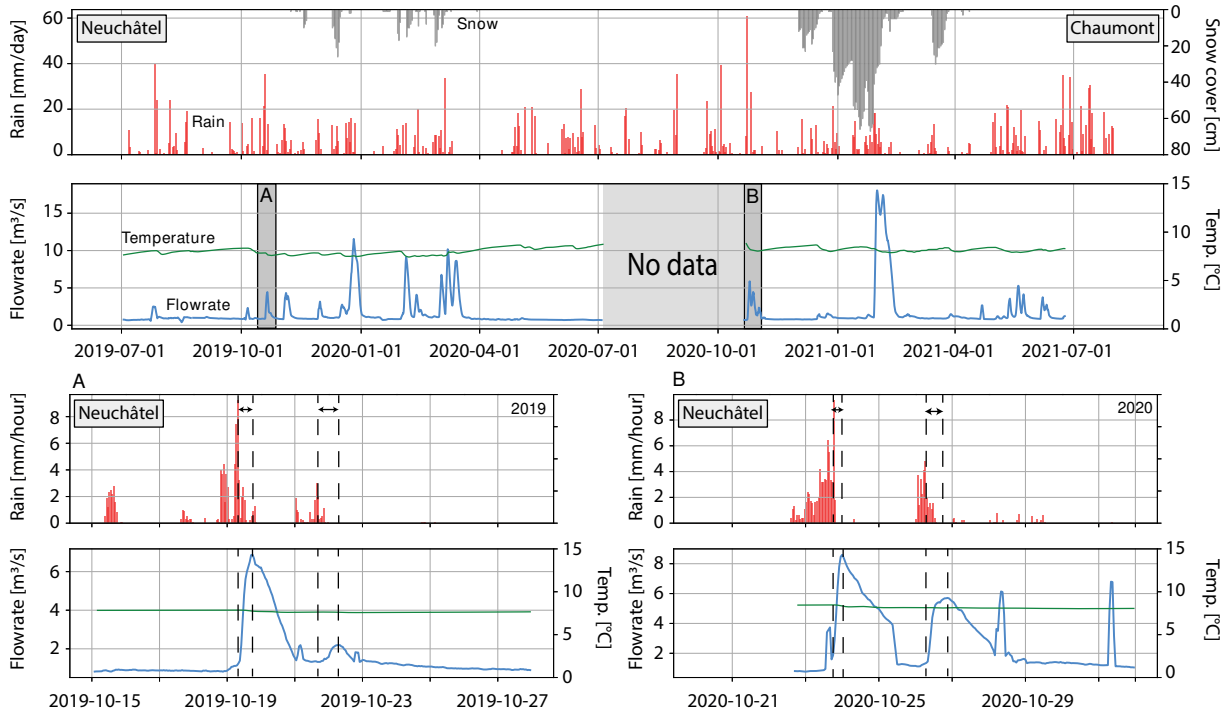


Figure 3.6: Hydrograph of the Serrière spring (blue line) from July 2019 to July 2021. Also shown, the rain (Neuchâtel station), the snow cover (Chaumont station) (red and grey bars, respectively) and the temperature (green line). Depredation to the logger explains the missing data at the end of summer 2020. Panels **A** and **B** show two distinct flood events in October 2019 and 2020, respectively. Dashed lines show the approximate time lag between the maximum rain intensity and its corresponding peak in the discharge curve.

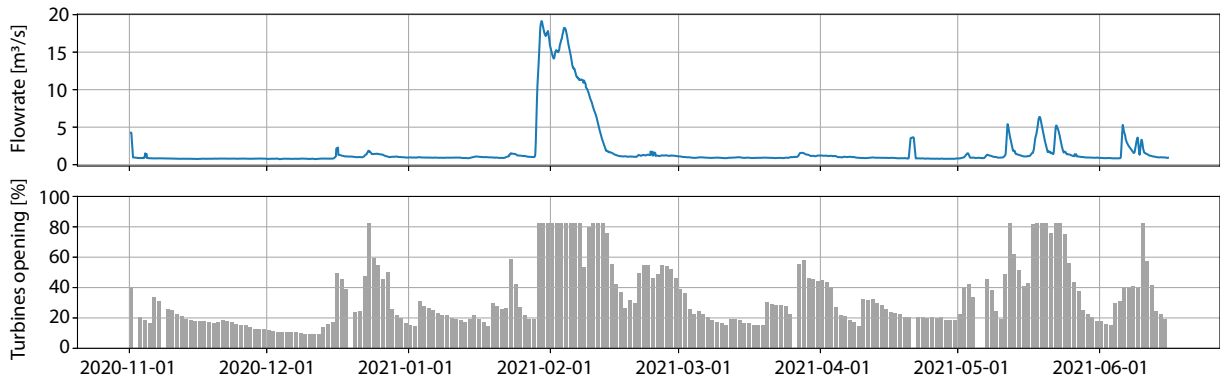


Figure 3.7: Flow rate of the Serrière spring from November 2020 to July 2021. Bottom panel shows the turbine's opening of the electrical power plant located downstream of the observation basin.

or the closing of the turbine's blades. A rapid level rise will activate the turbines, and the level will drop relatively rapidly in the basin. If the level keeps increasing, the turbine's blade will increase their opening, resulting in a rather stable water level in the basin. Hence, a given water level in the basin possibly reflects different flow rates, depending on the turbines activity. To evaluate the spring's regime and the flow rate variation, data of the turbines activity provided by the energy producer (Viteos) is plotted along with the flow rate for the time period 2020-

2021 (Figure 3.7). Considering both data sets, it is possible to determine whether the spring is undergoing high or low water conditions, but specific discharge values must be considered with caution. When reacting to a rainfall, the time lag between the maximum rain intensity and the maximum discharge is between 10 and 20 hours.

Dia

Figure 3.8 illustrates the Dia spring from July 2019 to July 2021. This spring activates only during high water periods, and reaches rapidly relatively important discharge values from 0 up to 5000 l/s. The activation of the spring corresponds to periods of relatively high waters, after precipitation and snow melt, especially in winter and at the beginning of the spring. The water temperature curve clearly indicates when the logger is immersed under water or when it is measuring air temperature. When the spring is not activated, the temperature fluctuates over the day, whereas when the logger is under water, the values stabilize around 9°C. During high water, the time lag between the maximum precipitation intensity and the maximum flow rate is around 6-8 hours. During periods of low waters, precipitations do not affect the spring discharge.

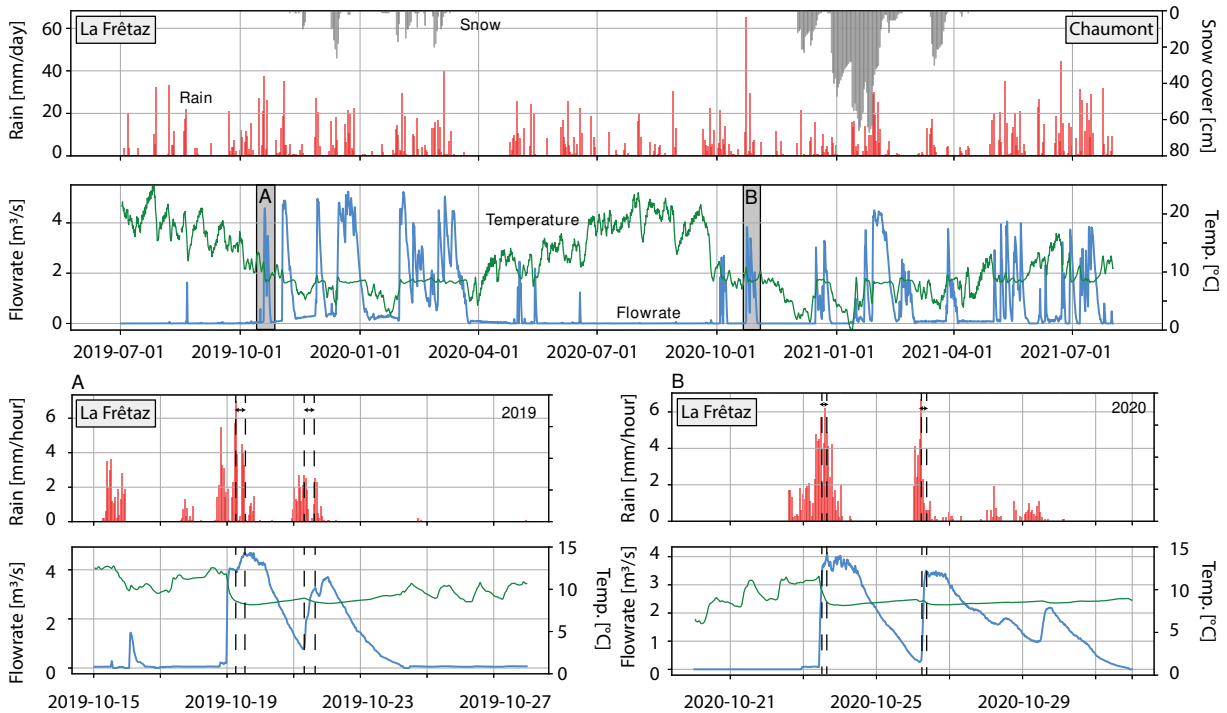


Figure 3.8: Hydrograph of the Dia temporary spring (blue line) from July 2019 to July 2021. Also shown, the rain (La Frêtaz station), the snow cover (Chaumont station) (red and grey bars, respectively) and the temperature (green line). When the Dia is not activated, the temperature curve shows the air temperature. Panels A and B show two distinct flood events in October 2019 and 2020, respectively. Dashed lines show the approximate time lag between the maximum rain intensity and its corresponding peak in the discharge curve.

Torrent

This spring is the top of a 130 m deep sinkhole. The water level is located at approximately -128 m below the surface during low flow conditions. The level logger is installed at the bottom of the sinkhole and measures the water level continuously (Huguenin, 2020). After heavy precipitation and/or rapid snow melt, the water level rises along the sinkhole, eventually reaching the surface and flowing down a temporary stream bed. Figure 3.9 illustrates the water level variation within this sinkhole from September 2019 to July 2021. Over the period of observation, the Torrent was activated only once in February 2021, when the water level reached +130 m, which is the total depth of the sinkhole. Nonetheless, several other important level variation occurred over the observation period, all related to either snow melt or important precipitation. The reaction time between the maximum rain intensity and the maximum water head is between 12 and 20 hours. The temperature of the stagnating water at the bottom of the Torrent sinkhole equilibrates with the air temperature of the sinkhole, around 7.5°C, while there is a positive pulse followed by a negative pulse every time the water level increases. The positive pulse is thought to be the pressure response of the increasing hydraulic head, where the water already present in the surrounding karstic conduits fills the sinkhole, while the negative pulse corresponds to the arrival of the colder rain water. This is supported by the amplitude of the negative pulse, which is greater during winter months.

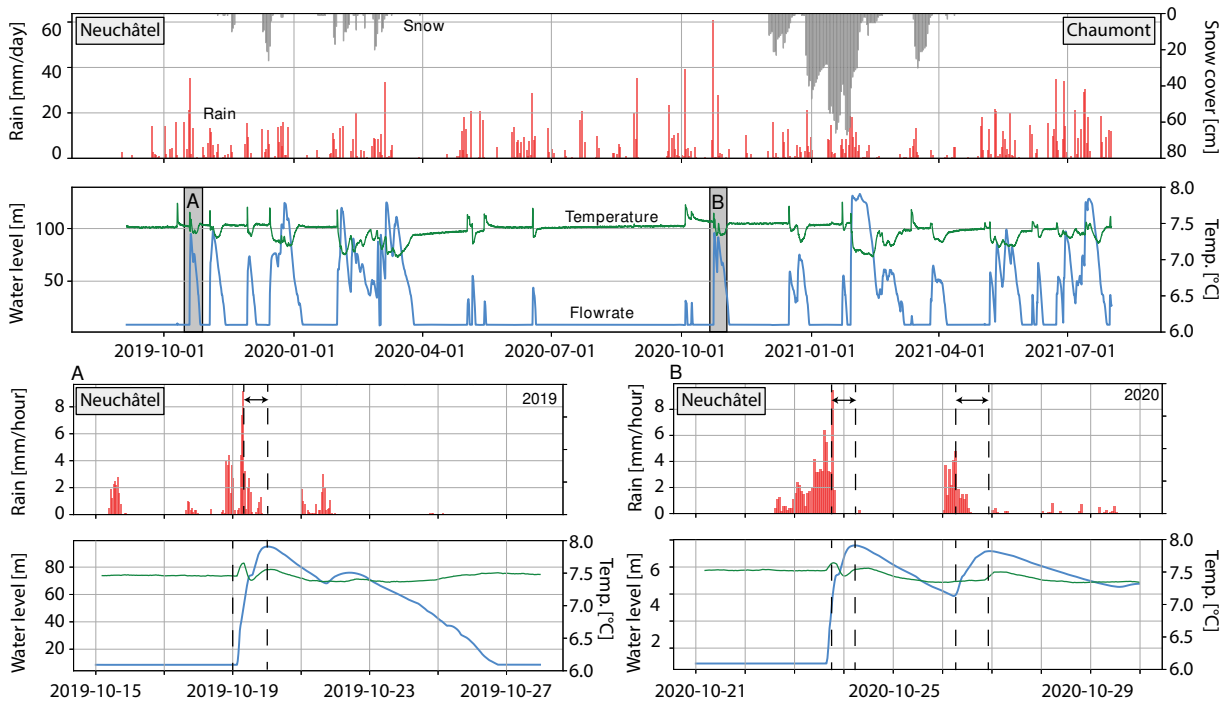


Figure 3.9: Water level in the Torrent sinkhole (blue line) from September 2019 to July 2021. Also shown, the rain (Neuchâtel station), the snow cover (Chaumont station) (red and grey bars, respectively), and the temperature (green line). Panels A and B show two distinct flood events in October 2019 and 2020, respectively. Dashed lines show the approximate time lag between the maximum rain intensity and its corresponding peak in the water level curve. Data from the Spéléo Club des Montagnes Neuchâteloise, retrieved from <https://www.hoellochforschung.ch/en>, in August 2021.

Discussion

The hydrographs represent the global response of the aquifer. In karstic areas, it represent the response of the low porosity matrix and the highly permeable conduits. This duality in hydraulic parameters (Király, 1994) results in the typical shape of karstic spring hydrographs, with a rapid increase and decrease in flow rate during and shortly after the flood (filling and emptying of the high permeable conduits) followed by a slower recession, resulting from the emptying of the matrix. This typical shape is found in the Raisse VD hydrograph (Figure 3.5). For the Serrière spring, the hydrograph resulting from the water level measurements shows a diversity in responses following similar precipitation events. This lack of consistency is due to the hydroelectric power plant installed at the spring. The Dia and the Torrent show rapid water level increase but a slower recession, relatively constant. Figure 3.10 shows normalized data of the 4 springs for comparison. For a same flood, the rising limb of the Raisse VD and the Dia are the steepest. The recession curve is steep for the Raisse VD shortly after the flood and flattens over time. The recession curves of the Dia and the Torrent are almost linear. For the flood of November 2019, the Dia and the Torrent go back to their pre-flood conditions after 12 days, while the flow rate at the Raisse VD after 12 days is still +20% greater than before the flood. Flow rate at the Raisse VD 20 days after the flood is still between +5-10% over the initial discharge value. In a general manner, hydrographs of floods for the different springs tend to indicate that the time lag between the maximum rain intensity and the peak in discharge is shorter for the Raisse VD and the Dia (6-15 hours) as for the Serrière and the Torrent (10-20 hours), indicating a possible higher karstification for the Raisse VD - Dia system.

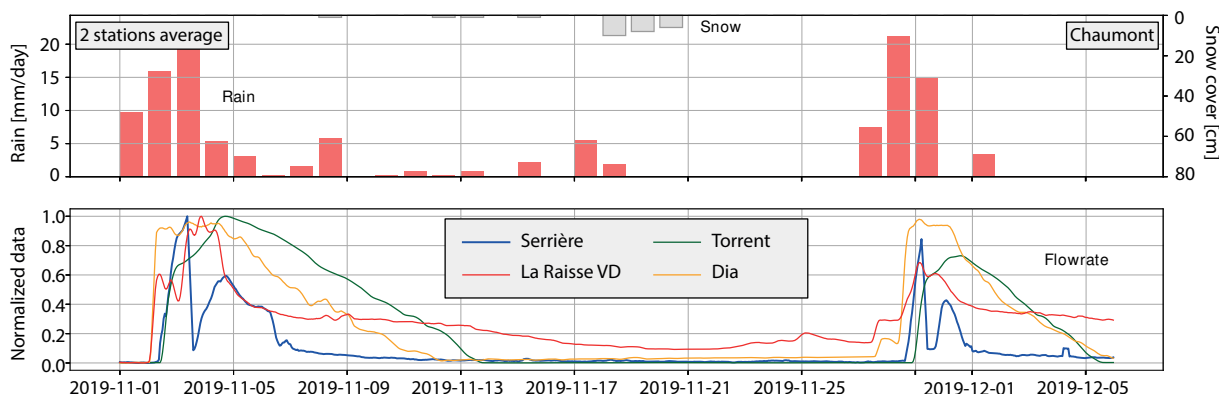


Figure 3.10: Top: Precipitation and snow cover, average of stations La Frêtaz and Neuchâtel for the rain, and data of the Chaumont station for the snow. Bottom: Normalized data of the 4 springs during the month of November 2019, showing 2 successive floods.

The activation of the two overflows, the Dia and the Torrent, are good indicators of the fluctuating hydraulic gradient as the water level necessarily reached at least their altitude while they were flowing. Hence the difference in altitude between the permanent spring and their overflow is a precious information regarding the increasing hydraulic pressure at depth. The Torrent sinkhole is of particular interest since data shows the variation of the hydraulic head within the sinkhole. Comparing the annual normalized data of each spring (Figure 3.11), shows

that during the period of observation, the 4 springs react to the same precipitation and/or snow melt event, and that the majority of the floods are of regional importance.

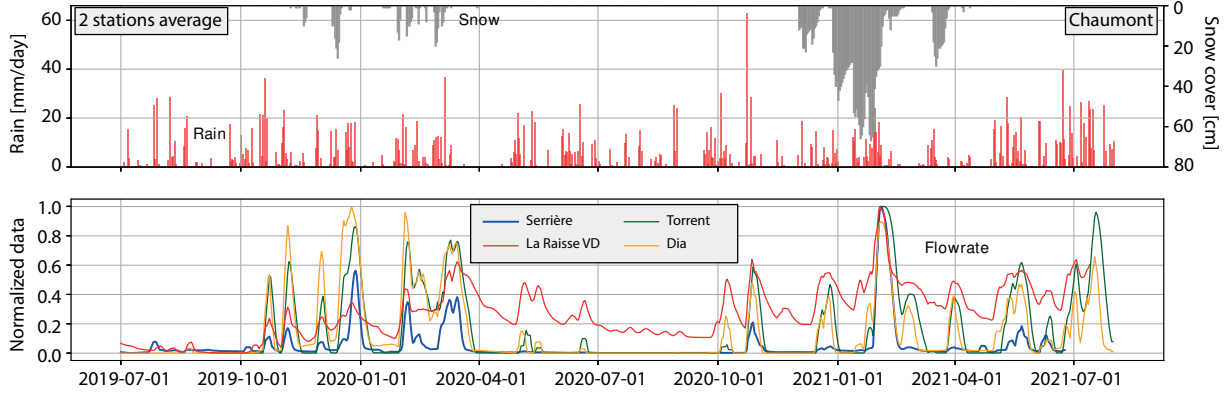


Figure 3.11: Top: Precipitation and snow cover, average of stations La Frêtaz and Neuchâtel for the rain, and data of the Chaumont station for the snow. Bottom: Normalized data of the 4 springs during the observation period from July 2019 to July 2021. Major floods are of regional importance and the majority of smaller floods are still visible on all hydrographs.

Seismic catalog

The continuous seismic monitoring of the study area resulted in a local seismic catalog containing 109 events identified as local earthquakes (LQ) and 20 events identified as local but uncertain because detected on very few stations (LU). Local magnitudes for the events identified as LQ range between -1.0 and 2.5. Hypocenters lie mostly between 0 and 4 km depth. The list of detected earthquakes can be found in Appendix E. Figure 3.12 illustrates the seismic catalog after 2 years of continuous seismic monitoring as well as the depth distribution and error. The central and southern part of Lake Neuchâtel, as well as the area located west of Neuchâtel and further north, seem to be the most active zones. From the collected data, the magnitude of completeness (M_c) is estimated to be between 0.5 and 1 (Figure 3.13). With a M_c of 0.5, the recurrence parameters a and b of the Gutenberg-Richter law are $a = 2.21$ and $b = 0.83$, for a M_c of 1.0, $a = 2.15$ and $b = 0.80$. Both scenarios are in relatively good agreement with values estimated for the area of interest in Wiemer et al. (2009) (tectonic and seismogenic zones 10 and 23, respectively, in Wiemer et al. (2009)).

Cross-correlation and Focal Mechanisms

The seismic signal of the 129 events were cross-correlated using the CORR program in SeisAn in order to identify events with similar waveforms (see Ottemöller et al. (2017) for description of the program). The input parameters for cross-correlation and the resulting cross-correlation coefficient are presented in Appendix G. These cross-correlation coefficient (ccc) were then used to detect groups of similar events using the clustering program in SeisAn (XCLUST). Pairs of events with $ccc = 0.7$ or higher were grouped into clusters (see output file of XCLUST in Appendix G). A total of 17 clusters regrouping 78 out of 129 events were detected (see map in Appendix G). Some events preliminary identified as LU show high correlation with events

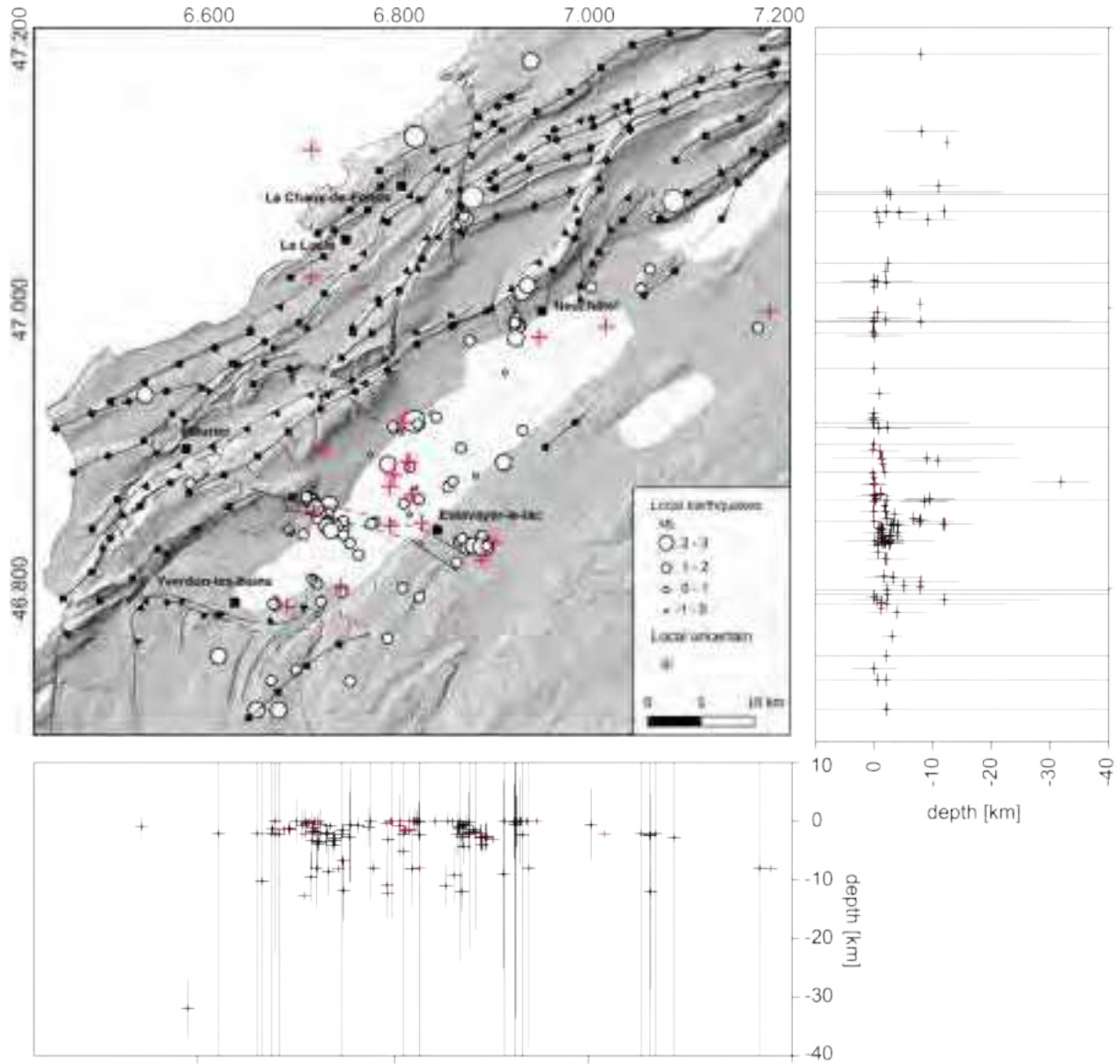


Figure 3.12: Maps of Local earthquakes epicenters and depth distribution and error (LQ, white dots and black cross), as well as the location of events identified as Local uncertain (LU, red cross).

identified as *LQ*, increasing the number of events identified as local earthquakes. As waveform similarity is caused by proximity in hypocenters and similar focal mechanisms, events belonging to the same cluster are used to estimate composite fault plane solutions using the first-motion polarity method. Figure 3.14 shows the estimated focal mechanisms of 5 clusters, determined using the HASH algorithm (Hardebeck and Shearer, 2002) implemented in SeisAn. Solutions were found using a 5 to 10 degree increment for the grid search and allowing between 1 and 5 polarity reading errors (see Appendix H). Focal spheres represented on the map show the 2 nodal planes, as well as the polarity distribution and the projection of the direction of the maximum compression and dilation axis, P and T, respectively.

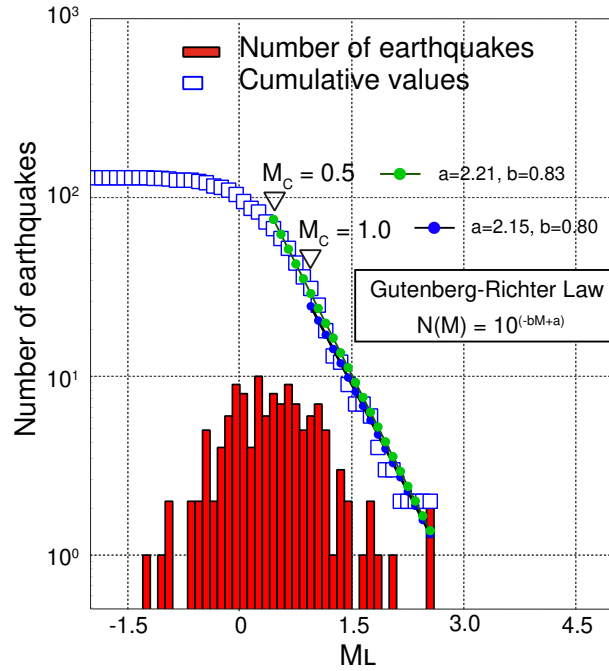


Figure 3.13: Gutenberg-Richter law's recurrence parameters a and b and magnitude of completeness

Relative relocation

Earthquakes have been relocated using the double-difference technique implemented in HypoDD (Waldhauser and Ellsworth, 2000). The double-difference technique assumes that for a pair of events with an inter-event distance relatively small compared to the event-station distance, the travel time difference from the source to a common station between two events is attributed to their spatial offset. Any structural bias is cancelled out since the ray path between the source and the station is similar for both events. The fundamental data to relatively relocate earthquake with HypoDD are travel time differences for pairs of event. These travel time differences can either be differences between the absolute time observations (dt_{ct}), or time differences inferred from waveform cross-correlation (dt_{cc}). The algorithm looks at the residuals between the observed or inferred travel times and the calculated travel times for a pair of events at a common station. The method of singular value decomposition (SVD) is employed to minimize the double-difference residuals for pairs of earthquakes at each station through weighted least squares. By changing hypocentral parameters (latitude, longitude, depth and origin time) HypoDD tries to find the best solution for the model to fit the data (see Waldhauser and Ellsworth (2000) for complete description of the technique). The maximum distance between pairs of events and the threshold for residuals must be set before running HypoDD. During iteration, the inter-event distance and the residuals will change, if these values overcome the thresholds for given pairs of events, they are disregarded for relocation. Figure 3.15 illustrates the result of the relocation of 43 events out of 105. Hypocenters of events in clusters 1 and 4 have been shifted closer together during iteration and are now more or less aligned on a N-S segment and a NW-SE segment, respectively. Hypocenters of events in cluster 5 and 6 converged into a single location (± 20 m). Hypocenters of cluster 16 were shifted along a NNW-SSE segment. Other isolated pairs

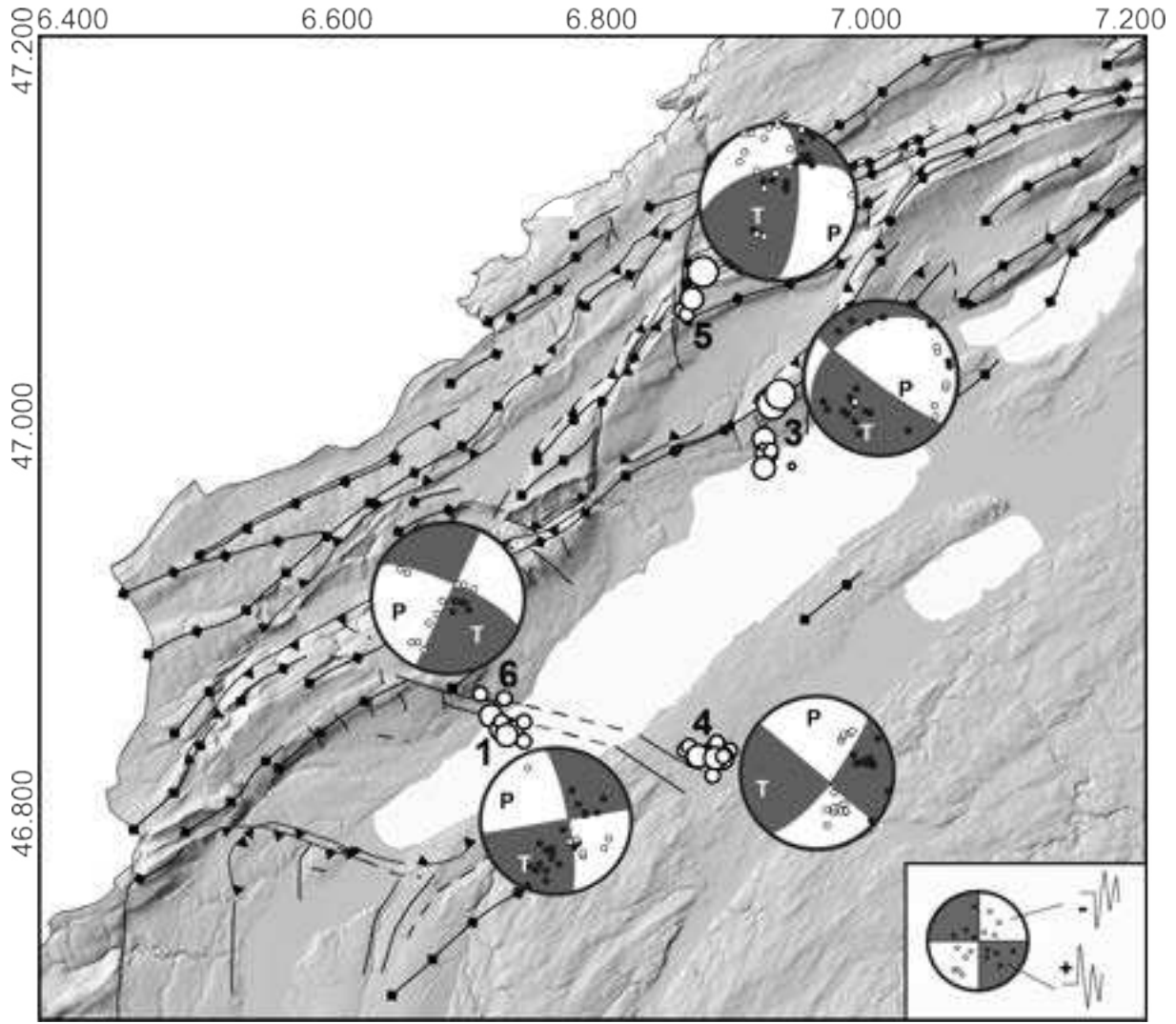


Figure 3.14: Estimation of focal mechanisms based on composite fault plane solution using events of same clusters (Clusters 1, 3-6). Also shown the distribution of the polarities reading ((-) white circle, (+) black circle) and the projection of the compression and dilation axis, P and T, respectively.

of events were shifted closer together such as events in clusters 2, 7, 10, 12 and 15. The depth variation of relocated events varies between 0.1 and 42 km, but with a majority of events having been relocated within ± 2 km (see Appendix [J](#)).

Template Matching

Template matching is a process that identifies earthquakes based on waveform similarity by scanning continuous seismic data with a template. This process detects earthquakes that could not be identified with the classical STA/LTA method, and increases the number of earthquakes of lower magnitude and eventually detects additional seismic sequences. We use the EQcorrscan Python package ([Chamberlain et al., 2018](#)), an open-source software using the matched filter method to detect signals of similar origin (waveform similarity). This software allows scanning data of multiple stations with multiple templates at the same time. We select events belonging to

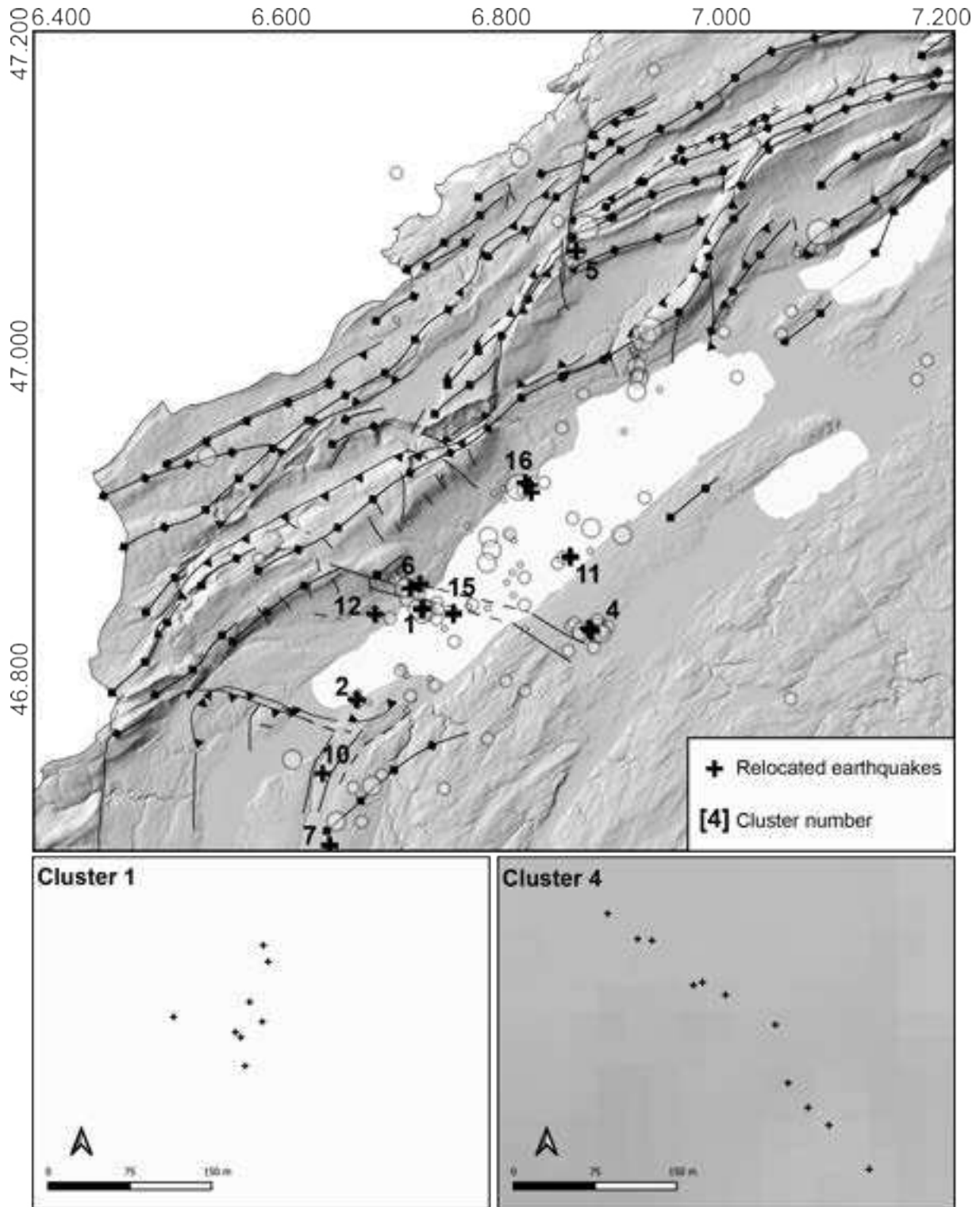


Figure 3.15: Map showing the relocated events using HypoDD (black cross). Panels on the right show the relocation of events in cluster 1 and 4. Numbers on the map refer to the clusters the relocated events belong to.

clusters and use their waveform on stations located in the vicinity of these clusters as templates. We create different groups of templates (called *Tribes* in EQcorrscan) for the different sub-networks (i.e. FER, BOU, LAN, EST). To create the templates, we down sample and filter the

data from 250 Hz to 125 Hz, and between 1 and 30 Hz using a 4th order Butterworth filter. We set a minimum signal-to-noise ratio of 5.0 for waveforms to be considered in the templates, and that the template contains at least 4 waveforms. The templates contain both P- and S-phases and start 0.1 seconds before the picks and last for 4 seconds. Continuous data is processed the same way as the template (filtered and down sampled). The normalized cross-correlation coefficient (CCC) between the template and continuous data is calculated for each individual channel. The addition of all CCC gives the cross-correlation coefficient sum (CCCS). If this value overcomes a given threshold, a detection is made. The threshold is set by multiplying the Median Absolute Deviation (MAD) of the CCCS of a day by a factor of 14. The multiplying factor is set such that the number of false detection is limited. Different values were tested and visual inspection of detections allowed to define 14 as the lowest possible multiplying factor to avoid too many false detections. These false detections are eliminated after final visual inspection. Data from the LAN sub-network is scanned by templates belonging to clusters 1, 2, 6, 10, 12, 13 and 15. Data from the FER sub-network is scanned by templates belonging to clusters 3, 5, 9 and 17. Data from the BOU sub-network is scanned by templates belonging to clusters 14 and 16 and data from the EST sub-network is scanned by templates belonging to clusters 4, 7 and 11. Events used for Template Matching and the *Tribes* they belong to are indicated in Appendix E. Events in Cluster 8 are not considered for Template Matching because located too far from the monitoring network. For the 2 years of data and the 56 templates, 106 supplementary events were identified (Appendix K shows the waveforms for which detections were made and 2 examples of detections). Figure 3.16 shows the detections and the clusters to which they belong. Clusters 1, 3, 4, 5, 11 and 16 show a relatively elevated number of detections, while clusters 6, 14 and 15 show very few or no supplementary detections. Results from Template Matching are in line with the preliminary catalog, where clusters containing many events show more detections than clusters containing fewer events, suggesting that some areas are more active than others. No significant additional sequence was identified as the detections mostly occur around the same period as their templates.

Discussion

One of the biggest challenges in seismology is to locate earthquakes accurately. The velocity model, the quality of the phase picking and the station distribution are among the parameters affecting the earthquake location and its quality. The choice of the velocity model used in our study is based on literature (i.e., Husen et al. (2003)) and is, to our knowledge and for the purpose of this study, sufficient to locate earthquakes accurately enough. The phase picking was performed by a single operator following the recommendations of Diehl et al. (2012) and a weight was attributed to each pick to consider its quality when locating the event (i.e. the highest the uncertainty the lower the weight). The station distribution affects the earthquake location by better constraining the solutions when stations are equally distributed around the epicenter. In our case, the station distribution depends on field constraints, where the presence of Lake Neuchâtel makes it impossible to install stations all around active areas. In order to evaluate the quality of the earthquake location performed in this study considering the used

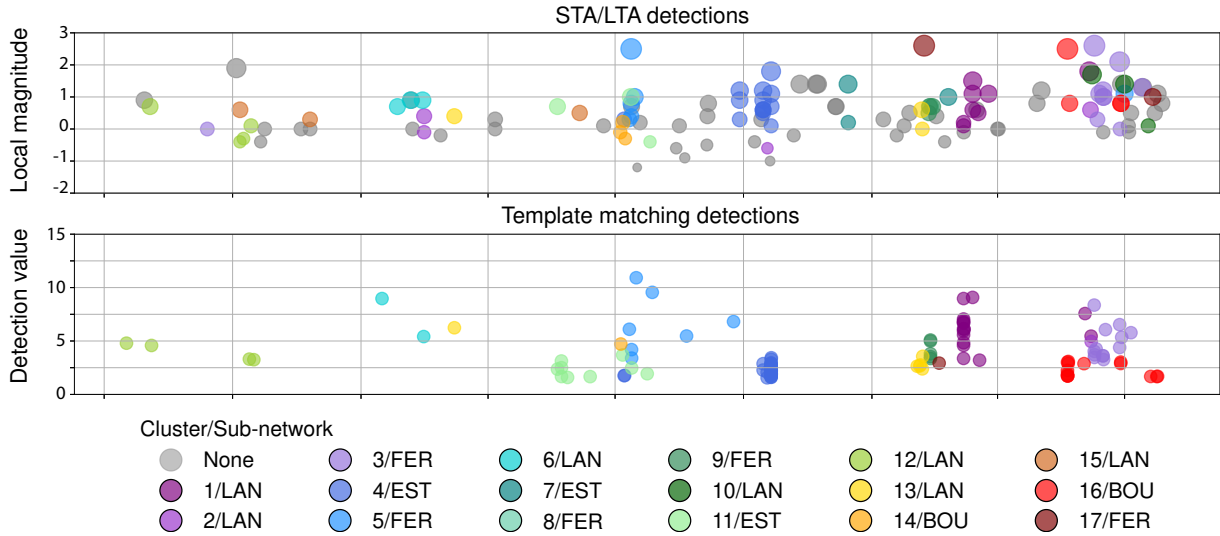


Figure 3.16: Earthquakes detected through STA/LTA and Template Matching techniques. Colors correspond to different cluster. For events detected through standard methods, the y-axis shows the Local magnitude, for events detected through Template Matching, the y-axis corresponds to the detection value (cross-correlation coefficient sum), which depends on the number of channel within the template. It does not reflect the magnitude of the event. Black contoured circles in the legend indicate that events in these clusters have been used for Template Matching.

velocity model, the phase-picking and the station distribution, the uncertainty for each event can be estimated using the following indicators (T. Kraft, personal communication):

1. Number of station used
2. Number of P and S arrivals used
3. Distance to closest station
4. Azimuthal gap
5. RMS of solution

Table 3.4 shows the threshold values to be reached in order for the event to be considered high-quality (i.e. low uncertainty in the location). Events that show values below these thresholds should be considered low-quality. For each event we attributed a score to the above mentioned indicator: 1 if the minimum value is reached and 0 if it is below. We allowed a score of 0.5 for indicators 1 and 2 if the number of stations and phases used are respectively 5 and 7, i.e. just below the threshold. An event is considered high-quality if its final score is 5 and low-quality if it is below 3. We attribute a medium-quality if it is between 3 and 5. Out of the 129 detected events, only 3 can be considered high-quality, events 92, 96 and 100 in Appendix F 78 are medium-quality and 48 are low-quality.

Although the majority of events have a medium to low-quality ranking, we still perform additional seismic investigations such as cross-correlation, composite fault plane solution and template matching, because the primary aim of this study is to determine whether seismicity

Parameters	Threshold value for high-quality ranking
Number of station used	6
Number of P/S arrivals used	8
Distance to closest station	$1.5 \times \text{depth [km]}$
Azimuthal gap	120 degrees
RMS of solution	0.2

Table 3.4: Threshold values for high-quality earthquake location ranking.

increases following important recharge and whether this data can be used to make a first estimation of the stress state. For the latter, determining fault plane solutions are essential but the reader must be aware that a fault plane solution resulting from low to medium-quality ranking events is a first estimation and should by no means be considered as the definitive solution for the fault's behaviour.

Focal mechanisms have been estimated using the composite fault plane solutions, using events belonging to same clusters to determine a single solution for a family of events. We use the absolute location and wave first arrival polarities for the estimation of fault plane solution but consider the output of HypoDD to identify the fault plane from the auxiliary plane, along with knowledge of local geological features and context. Figure 3.17 shows the fault plane solution (F1, F3-F6) and six additional fault plane solutions found in literature (Baer et al., 2007; Rabin et al., 2018; Lanza et al., 2022; Abednego, 2015) are also included. Strike, dip and rake of nodal planes are presented in Table 3.5 (see also Figure 3.18). To emphasis that the solutions found in this study are mainly resulting from low to medium-quality ranking earthquakes, we attribute a quality ranking to the solutions. We consider the highest score within the clusters but also the uncertainty in the strike of the fault and auxiliary plane (Appendix H). Since none of the fault plane solutions have uncertainties lower than 10° for both the fault plane and the auxiliary plane, all solutions are considered bad quality and are thus indicated in red on Figure 3.17.

Cluster 1 (F1) is located near an ESE-WNW right lateral strike-slip fault that could correspond to the approximately E-W nodal plane. After relocation, events in cluster 1 align on a small N-S segment that is also in agreement with the second nodal plane striking N-S. Both solutions are possible but the second integrates the position of the relocated events. Events in cluster 3 (F3) have not been relocated, the solution of the NW-SE striking plane with a normal slip component is considered as a fault plane associated to a probable NE-SW left lateral strike-slip fault. Cluster 4 (F4) shows a NW-SE nodal plane that corresponds to the supposed continuation of the La Lance fault on the southern shore of Lake Neuchâtel. Relocation of events in cluster 4 also show a NW-SE alignment and supports the solution of a NW-SE right-lateral strike-slip fault. Cluster 5 (F5) shows a N-S striking fault plane with a dominant left lateral strike-slip component associated to a reverse slip, this orientation is coherent with the Ferrière fault segment mapped in this area. Cluster 6 (F6) is located on the NW-SE La Lance fault but the estimated focal mechanisms does not match the right lateral sense of movement of this fault. As events in cluster 6 are located at depth $> 6\text{km}$, the fault plane solution for these events could correspond to a deformation structure within the basement, not necessarily linked with deformation in the sedimentary cover. For F1, F4 and F6, the stress-regime is almost purely strike-slip. For F3 and F5, important normal and reverse slip component are associated to the

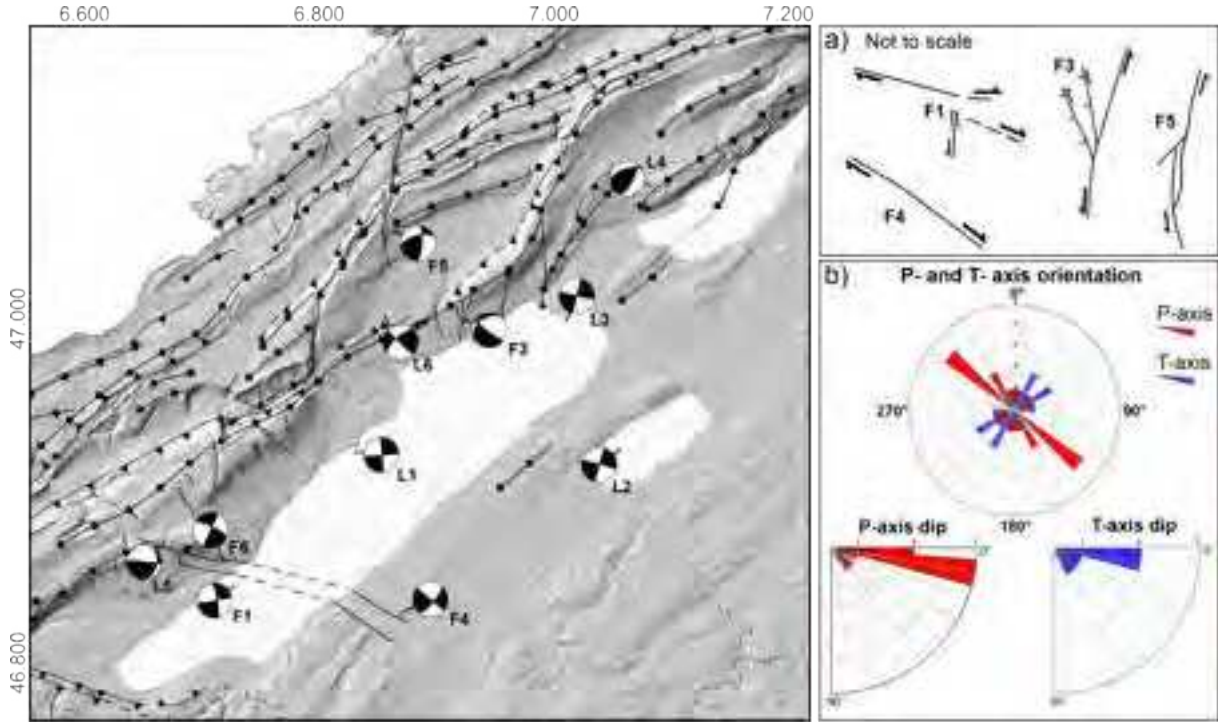


Figure 3.17: Compilation of focal mechanisms available for the study area. Letter **F** refers to focal mechanisms from this study. Letter **L** refers to focal mechanisms from literature (see text for references). a) Possible deformation structures corresponding to the focal mechanisms of F1, F3, F4 and F5. b) P- and T-axis orientation and dip values from the fault plane solutions.

strike-slip regime, respectively. Clusters 1, 3, 4 and 5 are considered to be in the sedimentary cover (depth between 0 and 4 km) while location of cluster 6 is uncertain (depth >6 km). Fault plane solutions from literature, L1 (Baer et al., 2007), L2 and L3 (Rabin et al., 2018) show almost a purely strike-slip regime. Location of the fault plane solution L1 corresponds to the epicenter of events in cluster 16 (Figure 3.15) that have been relocated on a NNW-SSE segment, possibly corresponding to the N-S nodal plane of the fault plane solution L1. L2 is most certainly located in the basement (depth of 30 km) while the location of L3 is uncertain with a depth of 7 km (Rabin et al., 2018). L4 (Lanza et al., 2022) and L6 (Abednego, 2015) show some reverse component and are both interpreted as being evidence for active thrusting in the Jura fold-and-thrust belt (Abednego, 2015; Lanza et al., 2022). L5 (Abednego, 2015) shows a vertical E-W nodal plane with a reverse right lateral component that could correspond to the La Lance fault. From all focal mechanisms, we observe that the dominant P- and T-axis orientations are NW-SE and NE-SW, respectively. Dipping of the P- and T-axis is limited and mostly lies between 0 and 20°.

3.5 Data set combination

To determine whether seismic events are triggered by significant recharge through heavy precipitation and/or seasonal snow melt infiltration, we combine the collected hydrogeologic and microseismic data. Figure 3.19 shows the time distribution of earthquakes together with the

Table 3.5: Strike, dip and rake of the nodal planes of the focal mechanisms shown in Figure 3.14 as well as P- and T-axis orientation and dip. Also indicated the approximate focal depth. FPS column refers to the name given to the different fault plane solutions. F: from this study, L: from literature (L1: Baer et al. (2007), L2 and L3: Rabin et al. (2018), L4: Lanza et al. (2022), L5 and L6: Abednego (2015)).

FPS	Depth	First nodal plane			Second nodal plane			P-axis		T-axis	
		Strike	Dip	Rake	Strike	Dip	Rake	Strike	Dip	Strike	Dip
F1	0-4 km	80	90	-165	350	75	0	306	11	214	11
F3	0-4 km	234	25	-160	126	82	-66	61	48	196	33
F4	0-4 km	310	90	160	40	70	0	357	14	263	14
F5	0-4 km	254	50	150	4	67	44	125	11	227	47
F6	6-8 km	25	90	-160	295	70	0	252	14	158	14
L1	2 km	355	86	-2	85	88	-176	310	4	220	1
L2	30 km	285	86	179	15	89	4	150	2	240	4
L3	7 km	13	90	0	213	90	180	148	0	58	0
L4	<5 km	250	40	128	24	60	63	134	11	246	64
L5	<5 km	15	25	10	276	86	115	345	36	210	44
L6	<5 km	135	85	15	228	60	6	185	17	87	24

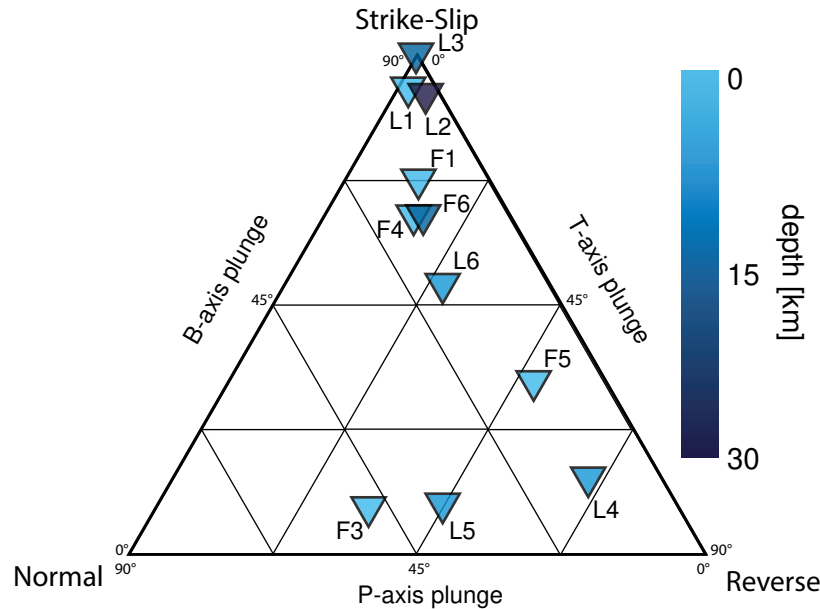


Figure 3.18: Focal mechanism ternary plot showing the fault plane solutions presented in Table 3.5. **L** refers to focal mechanisms from literature (see text for references) and **F** refers to focal mechanism from this study.

flow rate of the Raisse VD spring and precipitation data. Earthquakes are more frequent in the second half on the observation period, which corresponds to the time period between July 2020 and July 2021. The total amount of precipitation during this time period is higher than the cumulative precipitation between July 2019 and July 2020. We identify 6 periods with higher seismic activity referred to as periods S1 to S6 (Figure 3.19), occurring during different hydro-geological conditions. Events in S1 are associated with relatively low flow rates (200-1000 l/s), 2 of them (9 and 10, numbered earthquakes in Appendix E) occur shortly after a maximum flow rate of 2500 l/s. For S2, flow rates are mostly around 1800 l/s, events 15 and 22 occur shortly after two flow rate maxima of 3500 and 6000 l/s, respectively. The majority of events in S3

occur during the dry season, in July 2020. 3 events (25-27) occur before the beginning of the dry season, when the flow rate is around 1500 l/s. Events in S4 occur essentially mid-October when the flow rate is around 1500 l/s following a maximum flow rate of 3000 l/s but preceding a peak flow rate of 6300 l/s. Events in S5 occur during a period of high waters with an average flow rate of 2600 l/s. Events in the first half of February (77-85) occur following 2 successive flow rate maxima of 7500 and 6500 l/s, when the flow rate is still between 3000 and 4000 l/s. The maximum seismic rate in S6 occurs in June and July 2021. Similarly to S5, the average flow rate of the Raisse VD spring during earthquake occurrence is around 2600 l/s. Events 123-125 occurred between the 13 and 17 July 2021. At this time, Lake Neuchâtel reached its flooding limit with the highest water level ever recorded (430.57 m). Flow rate from the Raisse VD is not available for this time period because the level logger was flooded by the lake. However, the intensive lake level rise (+1 m) suggest an important flow rate at the spring.

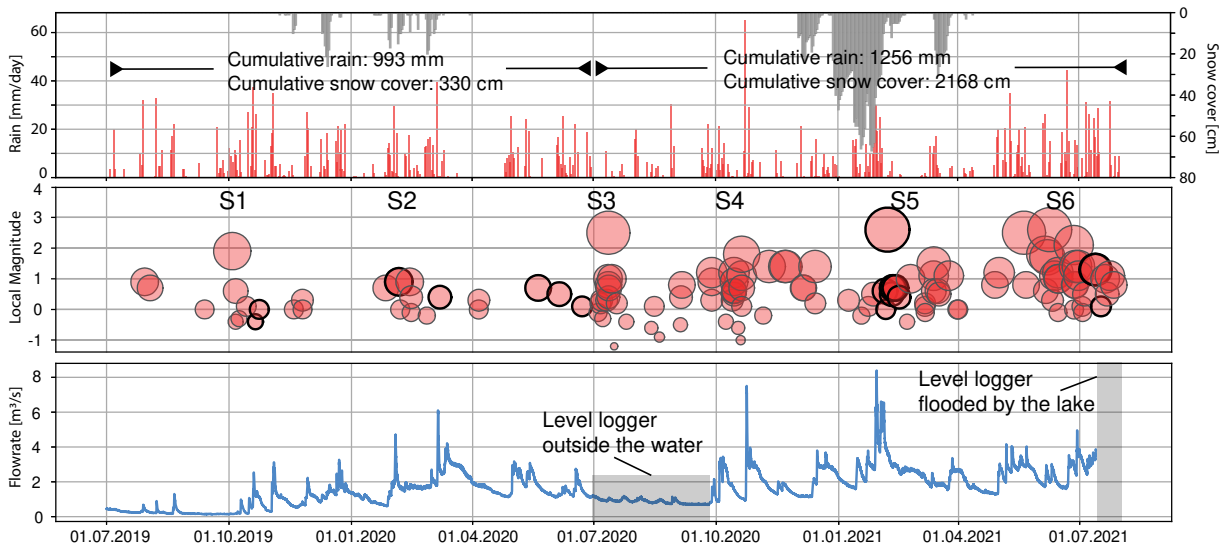


Figure 3.19: Combination of hydrogeologic and seismic data. Top of figure shows the precipitations and snow cover and their cumulative value for the years 2019-2020 and 2020-2021. In the middle, the earthquakes are represented as circles, the size corresponds to the magnitude. Circles highlighted in black are events mentioned in the text. At the bottom, the continuous flow rate of the Raisse VD spring. Letters S1-S6 refer to periods of higher seismic activity mentioned in the text.

3.6 Discussion and preliminary interpretations

To consistently evaluate whether the earthquakes are susceptible to be hydraulically triggered, we discriminate the seismic events between *dry* events and *wet* events according to the hydrogeologic conditions they occur under. We use the Torrent sinkhole water level to differentiate *wet* periods from *dry* periods, as it is a direct indication of the hydraulic head within the karstic network. We define *wet* conditions as periods for which the water level within the Torrent sinkhole (H_T) is equal or higher than 20 m. To account for a possible time lag between the moment the triggering pressure is reached and the earthquake, we also consider events as *wet* if the Torrent water level was ≥ 20 m within the 24h preceding the event. During the observation period, high

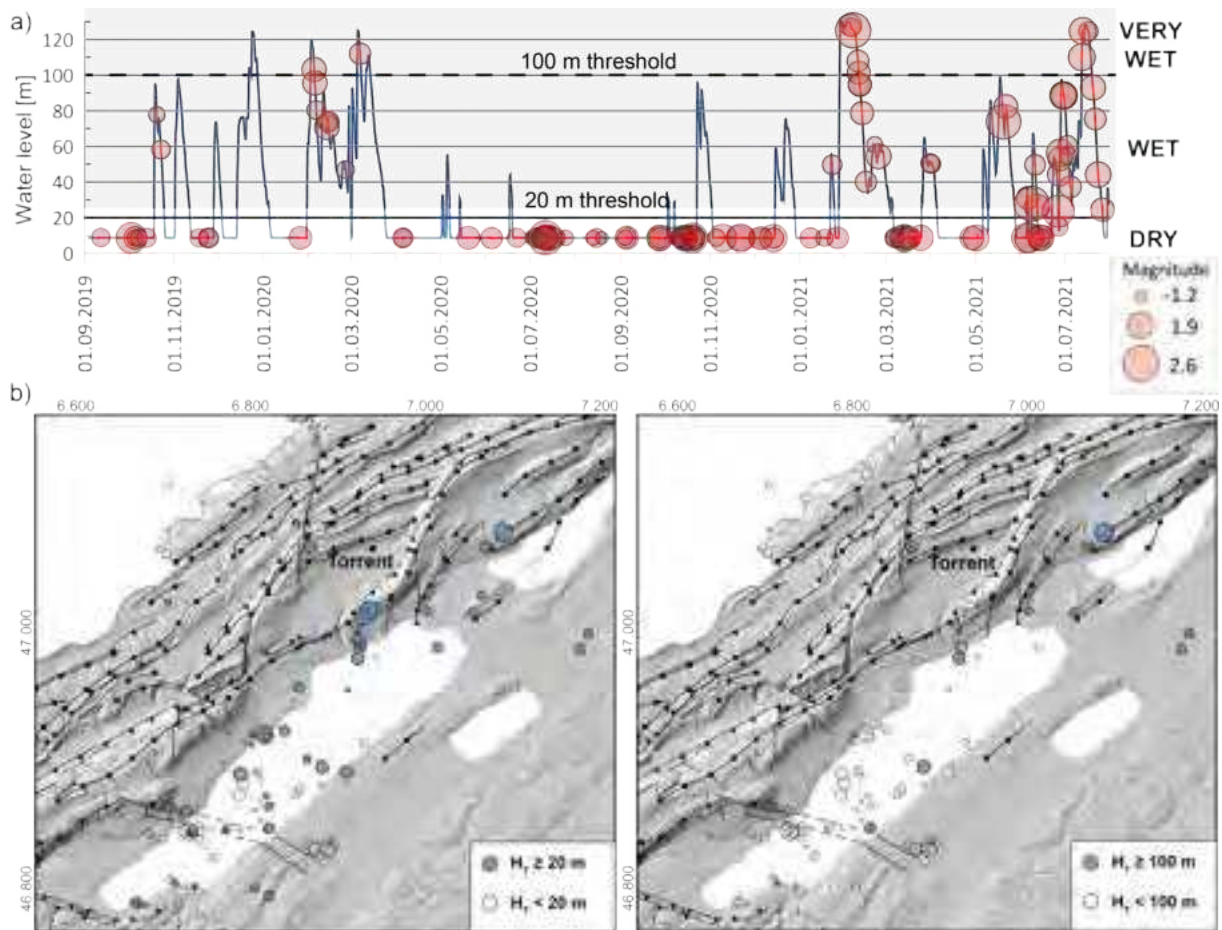


Figure 3.20: a) Continuous water level measurements within the Torrent sinkhole and earthquakes occurrence. Also indicated the two thresholds (solid and dashed lines). b) Maps showing the epicenters colored according to the water level within the Torrent at time of earthquakes. Left: threshold of 20 m, right: threshold of 100 m.

water conditions were reached 34.6% of the time. Out of the 129 identified events, 46 occurred during periods where the water level in the Torrent sinkhole was equal or higher than 20 m, and 5 occurred within 24 hours following high water conditions, corresponding to 39.5% of the total amount of earthquakes, suggesting a small increase in seismicity during high water conditions. To confirm this observation, historical data was analyzed by combining seismic events recorded by the Swiss National Institute for Seismology (SED) and flow rate of the Areuse, a major river originating from a karstic spring in the Neuchâtel area, since water levels of the Torrent are not available before September 2019. To evaluate what threshold value should be used to differentiate between low and high waters, the Areuse flow rates are plotted against the Torrent water levels for the time period both data sets are available (Figure 3.21). When the flow rate of the Areuse is $\geq 17 \text{ m}^3/\text{s}$, 95% of the associated water levels in the Torrent are $>20 \text{ m}$. Considering this percentage as satisfying, we set the threshold value between low and high waters to $17 \text{ m}^3/\text{s}$ at the Areuse river. From 2000 to 2022, 88 earthquakes with a maximum depth of approximately 5 km have been recorded by the SED over the larger area of interest. Among these events, 20 occurred while the Areuse had a discharge value of $17 \text{ m}^3/\text{s}$ or higher (Figure 3.22), meaning that 22.7% of the historical earthquakes can be considered as *wet*. Over the years 2000 to 2022, the

Areuse river was under high water conditions 17.7% of the time. This observation on historical data is a rough estimate since water level in the Torrent can be >20 m with flow rates at the Areuse spring as low as $8 \text{ m}^3/\text{s}$.

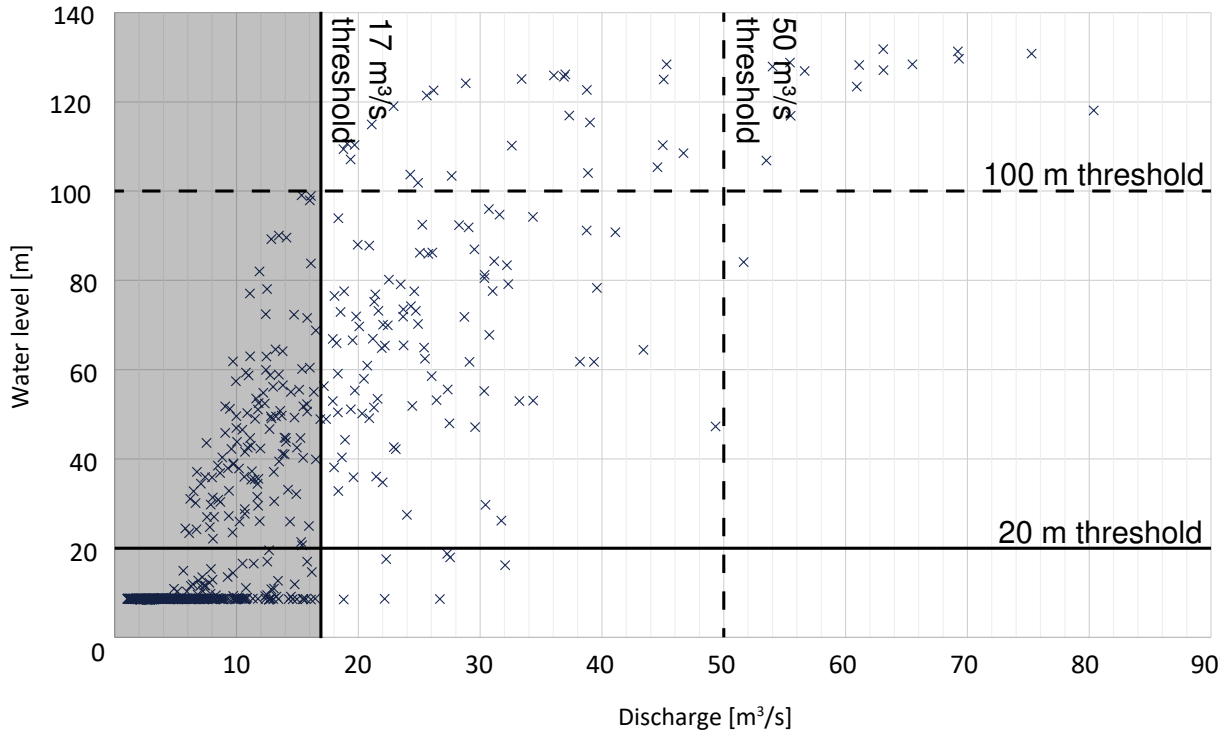


Figure 3.21: Water level within the Torrent sinkhole according to the flow rate at the Areuse spring. The average value for every day for which data is available for both observation sites have been plotted. The more the flow rate at the Areuse is high the more the water level within the sinkhole is elevated. Over $17 \text{ m}^3/\text{s}$, 95% of the data for the Torrent sinkhole is >20 m. Over $50 \text{ m}^3/\text{s}$ at the Areuse, 94% of the data for the Torrent sinkhole is >100 m.

From literature, we know that rain-trigger seismicity is not systematically occurring after every rainfall, but rather after exceptionally intense precipitations (Kraft et al., 2006b; Rigo et al., 2008). We differentiated our events into two categories, *wet* and *dry*, around an arbitrarily threshold value of $H_T \geq 20$ m. Although this value seems to indicate a recharge event (rainfall and/or snow melt infiltration), it does not necessarily indicate an extraordinary infiltration event. Thus, discriminating events between *wet* and *dry* periods is not sufficient and a supplementary differentiation should be made within the *wet* category, by discriminating the *very wet* events. We set the threshold over which we consider the period as *very wet* to $H_T \geq 100$ m. The water level within the Torrent sinkhole reaches or overcomes the threshold 5.2 % of the time over the monitoring period. Out of the 129 identified events 14 occur during very high water conditions or within 24 hours after, corresponding to 10.8 % of the total number of earthquakes. Figure 3.20 shows the spatial distribution of events considered as *wet* and *very wet*, information regarding the hydraulic conditions for each individual event can be found in Appendix E. Regarding the historical data, with a threshold between *wet* and *very wet* periods of $50 \text{ m}^3/\text{s}$ for the Areuse river (Figure 3.22), we get 4.5 % of earthquakes occurring during very high water conditions, representing 1.9 % of the total observation period. To evaluate whether these observations are

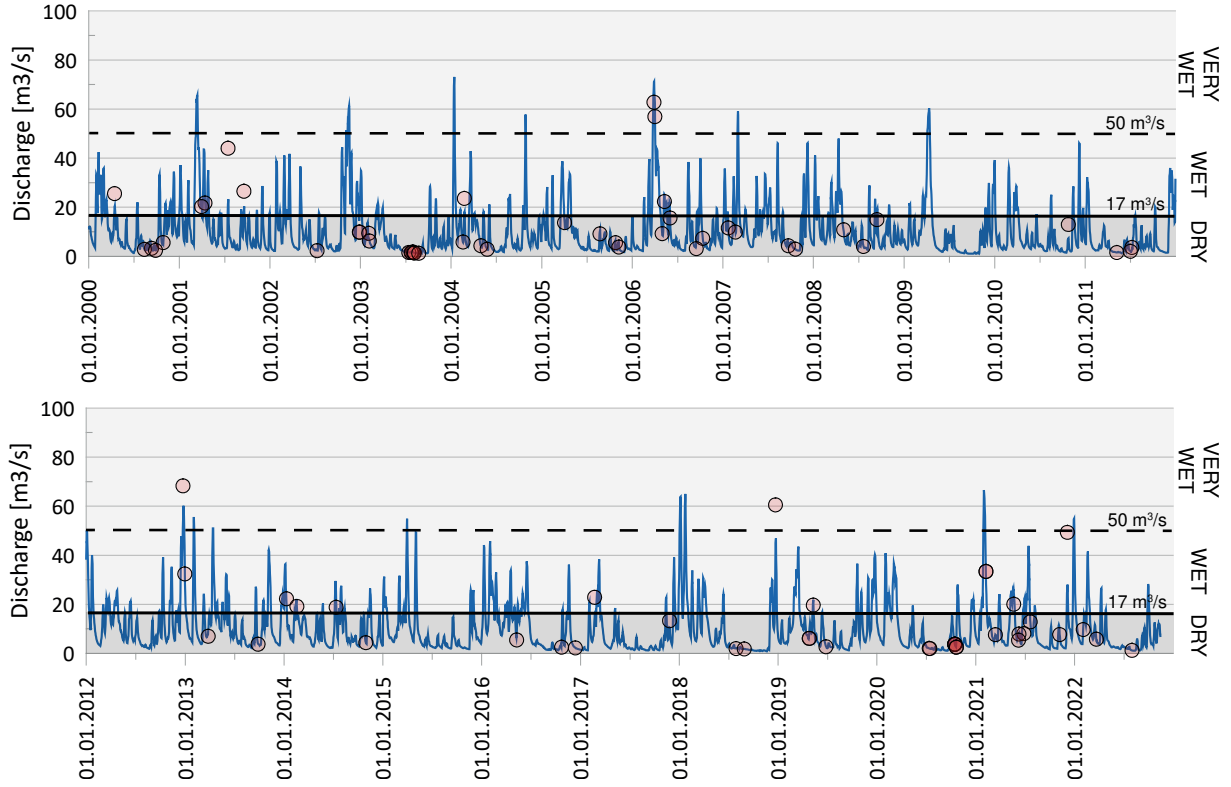


Figure 3.22: Continuous flow rate of the Areuse river from 2000 to 2011 (top) and from 2012 to 2022 (bottom). Solid line corresponds to the $17 \text{ m}^3/\text{s}$ flow rate threshold over which earthquakes are considered as *wet*. Dashed lined correspond to the threshold over which events are considered *very wet* (see text for further explanation). Red circles represent the historical events recorded by the Swiss Seismological Service.

statistically significant we perform Chi-squared test for both threshold values (see Appendix [L](#)). Two population are tested for independence, the first one is the number of hours during which the threshold is reached, the second is the number of hours during which the threshold is reached and for which there is an earthquake. The null hypothesis that there is no relation between high waters and earthquakes can be rejected if the $p\text{-value} \leq 0.05$. This value represents the probability of obtaining the observed outcome if the null hypothesis is true. For a threshold of $H_T \geq 20 \text{ m}$, $p\text{-value} = 0.16$ and for a threshold of $H_T \geq 100 \text{ m}$, $p\text{-value} = 0.0005$, which means that there is 16%, respectively 0.05 % of chance to obtain the observed outcome assuming there is no relation between high water conditions and earthquake occurrence.

3.7 Conclusion

Rain-triggered seismicity has been observed in different karstic regions such as Hochstaufen, Germany ([Kraft et al., 2006b](#)), Avignon France ([Rigo et al., 2008](#)) and Muotathal, Switzerland ([Husen et al., 2007](#)). Hypothesis regarding the mechanisms behind the triggering of seismicity in these regions have been presented in Chapter [2](#). It appears that the important hydraulic head variations within the karstic network generate large pore pressure changes at depth that can potentially trigger seismicity. Evaluating how much hydraulic head variation is needed to initiate

a fault rupture can inform on the proximity of the fault to failure. To develop this concept, field work was carried out in the Jura Mountains. Essentially composed of limestones layers from the Mesozoic, the Jura Mountains host numerous karstic galleries and major karstic springs and present a series of strike-slip faults that are potentially still undergoing deformation according to the general SE-NW horizontal principal stress orientation (Rabin et al., 2018). Results of our field work consists in continuous discharge values for the major karstic springs and in a seismic catalog for the area of interest. We observe that the majority of the seismic events occur in the southern part of the area of interest and have depths essentially between 0 and 5 km. Combining the hydrogeologic data with the seismic data reveals that earthquakes are approximately two times more likely to occur during high water conditions (10% of earthquakes occur during 5% of the time). Results from statistical tests show that there is most likely a relation between very high water conditions ($H_T \geq 100$ m) and earthquake occurrence. In Chapter 4, we focus on these *very wet* events and consider these earthquakes as rain-triggered. We apply the conceptual model presented in Chapter 2 and evaluate the criticality of the concerned faults.

Chapter 4

Hydrotectonics of the Neuchâtel Jura: insight from microseismic and hydrogeological data

4.1 Introduction

The interaction between water and tectonic forces, regrouped under the term of hydrotectonics, plays a major role in triggered or induced seismicity. Numerous studies detail the undeniable link between fluid circulation and the occurrence of earthquakes, as reported in the work of (Foulger et al., 2018) for the cases of human-induced seismicity. In addition, aftershock sequences following large (non-induced) earthquakes are thought to be fluid-driven (Miller et al., 2004; Gunatilake and Miller, 2022) and seasonal changes in groundwater recharge are reported as potential triggers for seismicity (Braunmiller et al., 2014; Saar and Manga, 2003; Bollinger et al., 2007; Bettinelli et al., 2008). If seismicity is sensitive to groundwater recharge fluctuations, i.e. following for example heavy rainfalls or important snow melt periods, the activated faults must be critically stressed, such that a relatively small stress perturbation initiates failure. In this context, the criticality of different fault zones located in the Neuchâtel Jura are evaluated by analysing data acquired between July 2019 and July 2021. Data comprises a seismic catalog, with earthquakes essentially located in the central and southern part of the study area as well as close to the city of Neuchâtel and further north. A few isolated events are found on the eastern side of the study area. In addition to the seismic catalog, continuous water level measurements within a sinkhole in the area of interest are used to determine whether the seismic events occur during low, high or very high groundwater conditions, respectively occurring during dry, wet or very wet periods. The aim is to assess the susceptibility of the concerned faults to fail during high groundwater conditions considering 3 key mechanisms: direct hydrogeologic connection, poro-elastic deformation and fluid pressure diffusion. These mechanisms influence fault stability by altering the effective normal stress, thereby influencing the shear strength of the fault. Identifying which mechanism is more likely to induce a rupture informs on the amount of stress change necessary to initiate a slip. In the subsequent chapter, a detailed description of the seismicity for 6 fault zones is performed using cross-correlation results from section 3.4.

Following this first step, focus is made on seismic events occurring during high to very high groundwater conditions. Analytical models are used to estimate pore pressure variations at time of rupture considering the 3 key mechanisms and using the varying water level within the sinkhole as input signal. Resulting pore pressure variation are then discussed and a preferential triggering mechanism is identified for each fault zone, together with an estimated critical pore pressure increase to be reached in order to observe seismicity. Results of the analysis are then used to discuss fault criticality and general aspects regarding the distribution of seismicity in the study area with regards to the geological and hydrogeological context.

4.2 Methods

Mechanisms potentially responsible for rain-triggered seismicity consist of: 1) Direct hydrogeologic connection, 2) Crustal loading and poro-elastic deformation and 3) Fluid pressure diffusion. The fault's stability is impacted through the reduction of the effective normal stress and/or the increasing vertical load, both influencing the shear strength of the fault. Pore pressure variations resulting from these mechanisms are calculated using equations presented in Chapter 2 and briefly re-discussed here for completeness.

Direct connection

The fault zone is in direct hydrogeologic connection with the karst conduits and the change in pore pressure directly relates to the increasing hydraulic head by the following relation:

$$\Delta p = \rho g \Delta h \quad (4.1)$$

where ρ is the water density, g the gravitational acceleration and Δh , the hydraulic head increase.

Poro-elastic deformation

The water filling the karst conduits acts as a supplementary load, inducing a poro-elastic deformation on an saturated underlying fault. The response will be an increasing pore pressure in response to the vertical loading. We use the Boussinesq's solution as expressed in Poulos and Davis (1974) to evaluate the impact of an increasing load P_a on the vertical stress beneath the center of a karstic network :

$$\Delta \sigma_z = P_a \left[1 - \frac{1}{[(R/z)^2 + 1]^{3/2}} \right] \quad (4.2)$$

where R is the radius of the circular area, and can be seen as the catchment area of the karstic network and z is the depth. This solution is then used to calculate the change in pore-pressure resulting from the change in vertical stress (Eq. (4.4)).

Under undrained conditions, the impact of stress changes on pore pressure is expressed as follows (Rice and Cleary, 1976):

$$\Delta p = B \frac{\Delta \sigma_{kk}}{3} \quad (4.3)$$

where $\Delta\sigma_{kk}/3$ corresponds to $(\Delta\sigma_x + \Delta\sigma_y + \Delta\sigma_z)/3$ and is the average compressive stress. Assuming that stress changes only in the vertical direction, then Eq. (4.3) becomes :

$$\Delta p = \frac{B(1 + \nu_u)}{3(1 - \nu_u)} \Delta\sigma_z \quad (4.4)$$

where B is the Skempton coefficient and ν_u the undrained Poisson's ratio.

Fluid pressure diffusion

There is no direct hydrogeologic connection between the karstic network and the fault zone. The pressure at the bottom of the karstic network (Eq. (4.1)) resulting from the hydraulic head increase diffuses through a permeable crust towards depth. Considering a pressure step increase at time $t = t_i$ the simplified 1-D diffusion problem has the solution :

$$p(z, t) = P_a H(t - t_i) \operatorname{erfc} \left(\frac{z}{2\sqrt{D(t - t_i)}} \right) \quad (4.5)$$

with the Heaviside function $H(t - t_i) = 1$ for $t \geq t_i$ and $H(t - t_i) = 0$ for $t < t_i$. While we considered a linear decrease over a time dt for the analytical model in Chapter 2, we here consider a pressure step increase of P_a followed by a pressure step decrease after a time dt . The solution reads as follows (principle of superposition, see Appendix A):

$$p(z, t) = P_a H(t - t_i) \operatorname{erfc} \left(\frac{z}{2\sqrt{D(t - t_i)}} \right) - P_a H(t - (t_i + dt)) \operatorname{erfc} \left(\frac{z}{2\sqrt{D(t - (t_i + dt))}} \right) \quad (4.6)$$

For varying pressures at the bottom of the karstic network resulting from varying hydraulic heads, we construct a succession of these pressure step increases and decreases that fits best the pressure history (Figure 4.1). Pore pressure variation at depth z and time t is simply the sum of these different steps :

$$p(z, t) = \sum_{i=0}^N P_i H(t - t_i) \operatorname{erfc} \left(\frac{z}{2\sqrt{D(t - t_i)}} \right) - P_i H(t - (t_i + dt)) \operatorname{erfc} \left(\frac{z}{2\sqrt{D(t - (t_i + dt))}} \right) \quad (4.7)$$

where P_i is the pressure at the bottom of the karstic network at time t_i . and N the number of steps.

4.3 Data

Assessing the criticality of the fault zones around lake Neuchâtel with regards to groundwater conditions implies the use of 2 distinct data sets. The first one is the seismic catalog presented in Chapter 3, containing the 129 events detected through standard methods, as well as the 106 events detected through template matching (see appendix E and K). While the absolute latitude

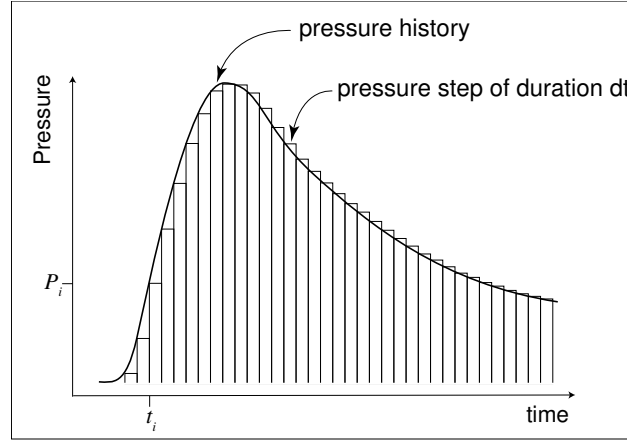


Figure 4.1: Schematic illustration of the pressure steps to re-create the pressure history.

and longitude are used for the detailed description, depth resulting from the hypoDD relocation process (see section 3.4) are considered when available, if the error in the relocated depth is within 2 km of the absolute depth (see Appendix J). The second data set consists in groundwater level measurement within a karstic sinkhole (here after referred to as the Torrent sinkhole), where the water level is stable during dry periods and rapidly increases following heavy rainfalls or snow melt periods (up to 120 m of fluctuation). As it has been observed that major floods are common to the entire study area (Section 3.4), the Torrent sinkhole water level is used as indicator for the hydrogeologic conditions (low, high or very high groundwater levels) for all zones of interest. Identification of events that are potentially triggered by an increasing groundwater level (referred to as rain-triggered events) is based on literature (Kraft et al., 2006b; Rigo et al., 2008; Hainzl et al., 2006) and conclusions from Chapter 2. We define rain-triggered events as earthquakes occurring within 48 hours following high to very high groundwater conditions, where the groundwater conditions are inferred from water level in the Torrent sinkhole. High to very high groundwater conditions occur when the water level fluctuation in the Torrent sinkhole is ≥ 90 m. Seismicity occurring within 48 hours after the groundwater level fluctuation reached this value are considered as rain-triggered. A detailed description of the seismicity is given for 6 fault zones: La Lance fault, Lessy fault, Neuchâtel and Monruz Faults, La Ferrière Fault, Saint-Blaise and La Neuveville villages (Figures 4.2 and 4.3), followed by the identification of rain-triggered events.

La Lance Fault

La Lance fault is an approximately 1 km-wide WNW-ESE dextral strike-slip fault extending from the Mont Aubert anticline towards the southern shore of Lake Neuchâtel (Figure 4.2) (Sommaruga, 1997; Gorin et al., 2003; Ndiaye et al., 2014). Figure 4.4 shows the seismic events associated to the La Lance fault. Over the 2 years of monitoring, it appears that the system is regularly activated. The different clusters and isolated events indicate that fault's activity is not restricted to a single source location and that different zones are activated over the monitoring period. From July 2019, seismicity starts with events in Cluster 12, located west of the village of Concise and probably located in the sedimentary cover with depth ranging between 0 and

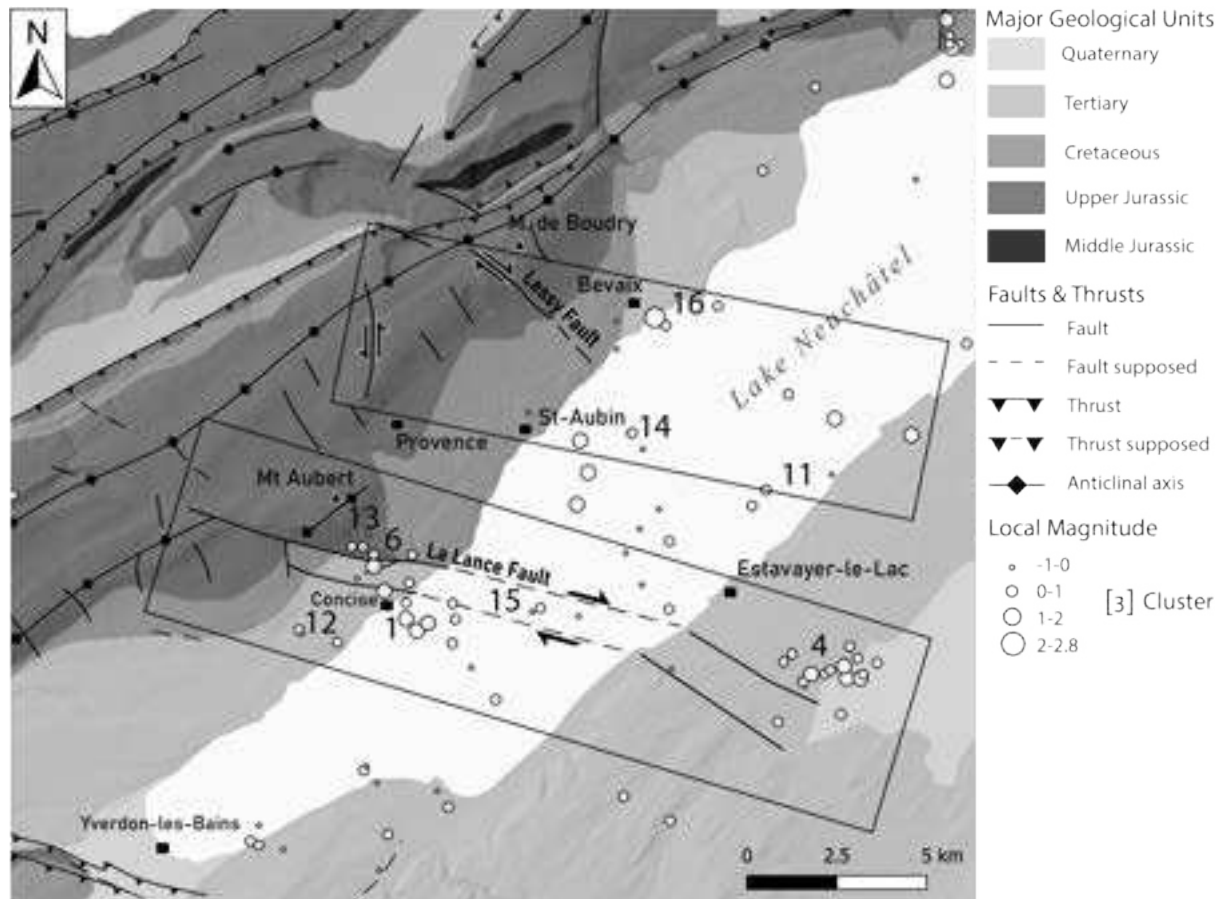


Figure 4.2: Tectonic sketch of the La Lance and Lessy Faults area. Seismic events considered for the zones of interest are inside the boxed areas. Background map compiled after the Geological and Tectonic map from the Federal Office of Topography, swisstopo.

1.4 km. This source seems to be active until October 2019, when a second source activates further East and deeper, with events in Cluster 15 (depth between 2-11 km, indicating large uncertainties in earthquake location) and even more to the East with 2 isolated events located between 0 and 2 km depth in November 2019. In January and February 2020, seismic events belonging to Cluster 6 occur North of the village of Concise, at depth between 6 and 10 km, followed rapidly by events in Cluster 13 with similar epicenters but with much shallower depth (0-0.3 km). Between April and June 2020, seismic activity consists in 1 isolated event located on the southern shore of Lake Neuchâtel at relatively shallow depth and 1 event belonging to Cluster 15. Except for an isolated event in August 2020 located near Clusters 6 and 13, seismicity seems to migrate towards the south with events in Cluster 4. A first event occurs in July 2020 followed by a swarm of 31 events in September and October 2020 in the region of Estavayer-le-Lac, with an average depth around 2 km. Following this relatively intense seismic activity on the southern shore of Lake Neuchâtel, no more activity seems to be directly related to the La Lance fault until February 2021, where the source origin of events in Cluster 13 is reactivated with the occurrence of 7 earthquakes in a relatively short amount of time, as well as the occurrence of an isolated event located below the lake, west of Estavayer-le-Lac. A month later, located southeast of the village of Concise, a swarm-like sequence takes place at an average depth of 2.7 km, with the occurrence of events belonging to Cluster 1. Between the end of March and June 2021, 2 isolated

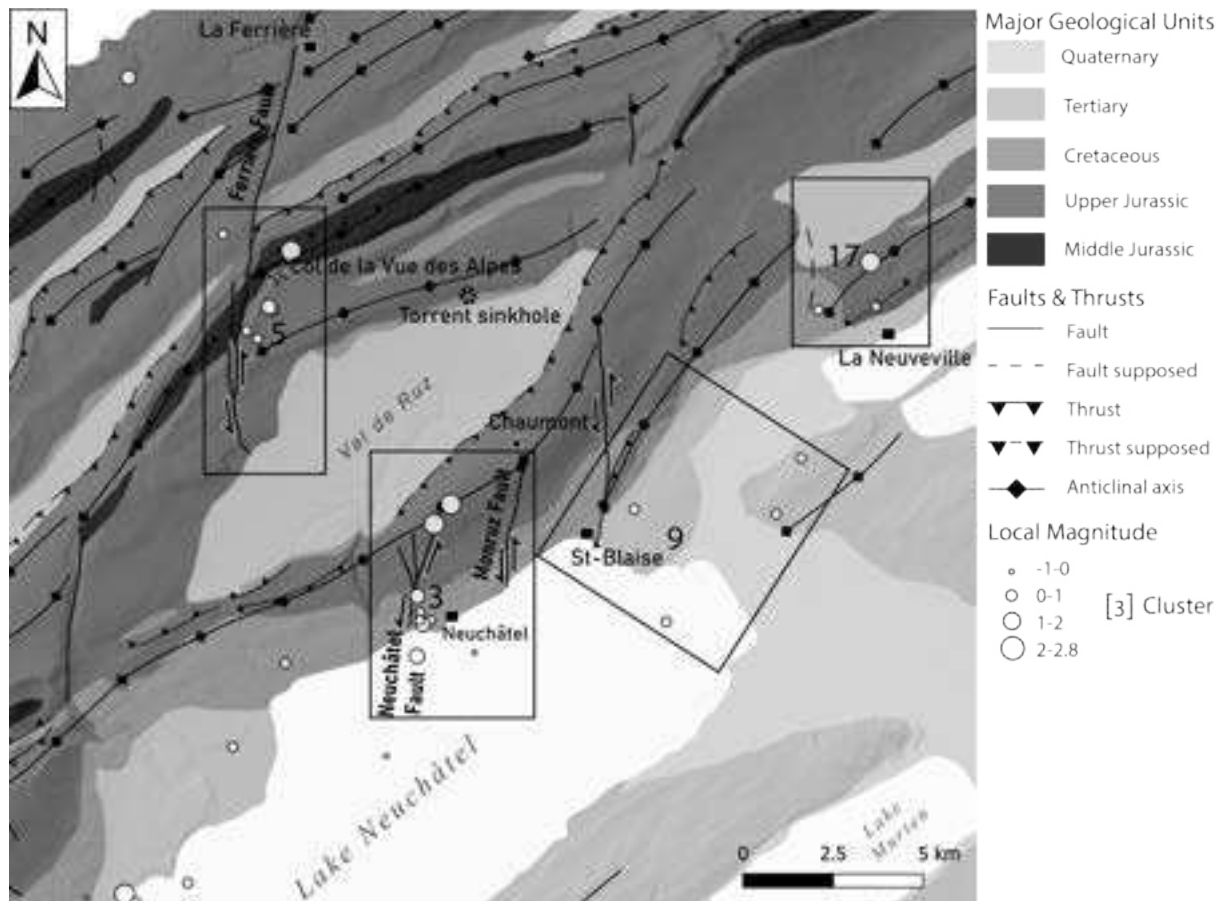


Figure 4.3: Tectonic sketch of the area comprising the Ferrière Fault, the Neuchâtel and Monruz Faults as well as the St-Blaise and La Neuveville areas. Seismic events considered for the zones of interest are inside the boxed areas on the map. Background map compiled after the geological and tectonic map from the Federal Office of Topography, swisstopo.

events take place East and North of the village of Concise, before the source origin of Cluster 1 is briefly reactivated in June 2021. This marks the last activity of the La Lance fault system over the 2 years of seismic monitoring.

Several events occur during periods of relatively high water levels within the Torrent sinkhole. Events in clusters 4, 12 and 15 occur before floods and are hence not considered as rain-triggered. The source origin of cluster 6 is first activated during a low water period in January 2020 and goes on while the Torrent water level increases in February and March 2020, when a second cluster, cluster 13 is activated shortly after a water level maximum, as well as an isolated event west of Estavayer-le-Lac. Cluster 13 is reactivated in February 2021. A month later, the source origin of cluster 1 is activated, relatively shortly after the water level drops.

Lessy Fault

The Lessy Fault is a NW-SE dextral strike-slip fault associated to a sinistral N-S strike-slip fault segment cross cutting the southern flank and top part of the anticline of the Montagne de Boudry, respectively (Figure 4.2) (Sommaruga, 1997). Although the southern continuation of the Lessy fault is not confirmed by surface geology (no clear offset), seismic activity is observed

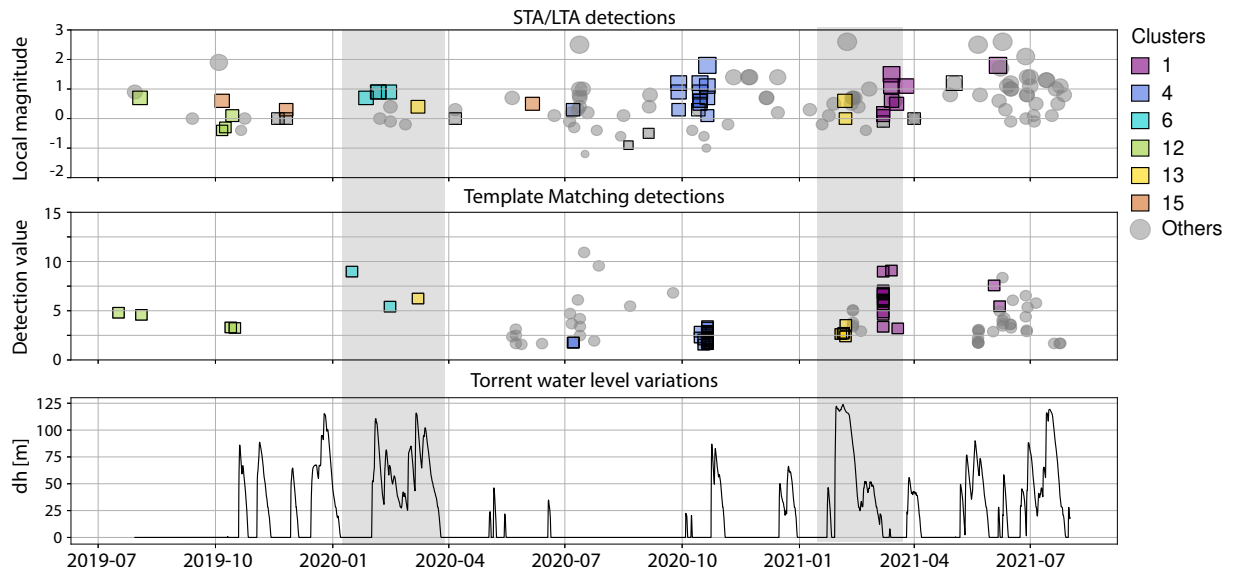


Figure 4.4: Time distribution of seismicity over July 2019 and July 2021. Black contoured squares correspond to events occurring near the la Lance Fault and discussed in the text. Colors correspond to clusters. Shaded areas refer to periods considered for the analytical modelling.

on the southern tip of the fault and off the shore of the lake, SE of Bevaix up to the other side of the lake. Events located around Saint-Aubin, are also considered in this zone as well as an isolated event near the associated sinistral fault, in the vicinity of Provence. Seismic activity begins in October 2019 (Fig 4.5) with the occurrence of a single event located near Bevaix at a relatively shallow depth. Until May 2020, no other seismic events associated to this region are recorded. End of May, a first event of cluster 11 is detected, followed by 2 additional events in July 2020. In June 2020, events belonging to cluster 14 and located off the coast of Saint-Aubin occur at an average depth of 1.5 km, followed in July, August and October by 3 isolated events, located respectively in Saint-Aubin, Provence and Bevaix. Until April 2021, no other seismic events are recorded. In May 2021, events detected through template matching seem to indicate a small sequence of seismic activity belonging to cluster 16. This sequence is active until the end of the July 2021. 3 isolated events, located close to the southern shore of the lake, and suggesting a SE-NW fault segments, occur from the beginning of July up to the end of the month, when the monitoring period ends.

Events belonging to clusters 11 and 14 as well as isolated events occurring between July and October 2020 happen under low water conditions and are not considered as rain-triggered. On the contrary, the first earthquake recorded in this area in October 2019 as well as events in cluster 16 occur during flood periods. One isolated event on the southern shore of the lake occurring on the 14th of July also happens during high water conditions.

Neuchâtel

On the southern flank of the Chaumont anticline, 2 sub-parallel fault segments striking NNE-SSW show an apparent sinistral offset (Sommaruga, 1997), the Neuchâtel fault and the Monruz fault (Figure 4.3). The Neuchâtel fault shows also 2 additional fault segments striking N-

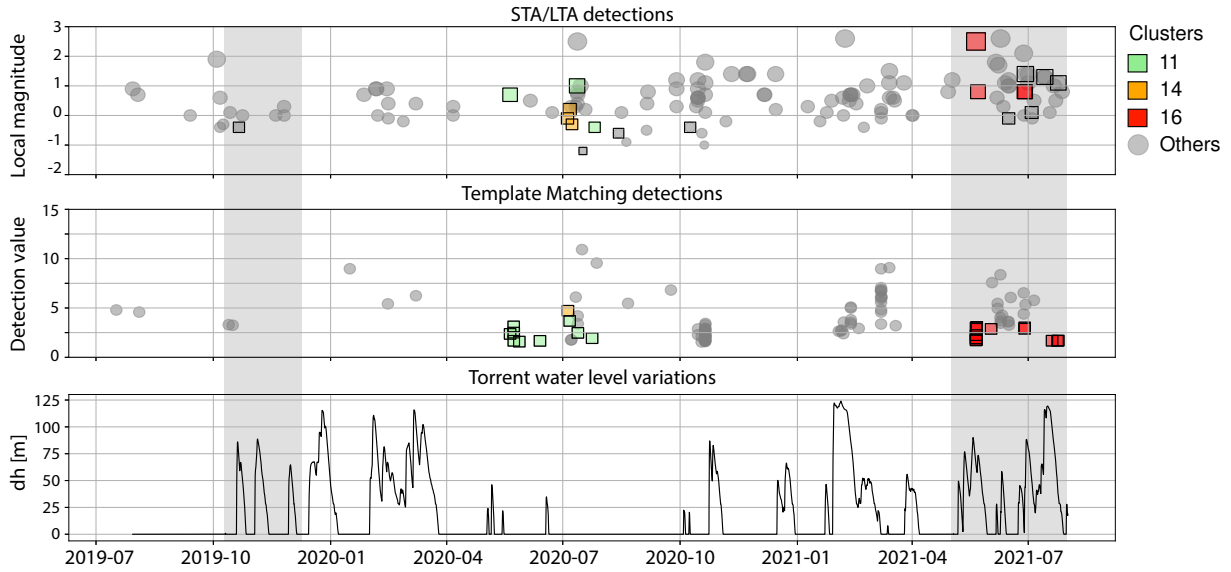


Figure 4.5: Time distribution of seismicity over July 2019 and July 2021. Black contoured squares correspond to events occurring near the Lessy Fault and discussed in the text. Colors correspond to clusters. Shaded areas refer to periods considered for the analytical modelling.

S. Events occurring in the vicinity of these 2 faults belong to the same cluster (cluster 3). The source origin of this cluster is activated once in September 2019 with a single event, and reactivated in June 2021 until the end of the monitoring period, with a peak activity between the 9th and the 15th of June.

While the first event is isolated and occurs during low water conditions in September 2019, events starting in June 2021 are numerous and occur during an overall relatively wet period (May-July 2021). However, as seismicity starts just before the maximum water level variation of June with a 2.6 ML earthquake, these events are not considered for the analytical modelling.

La Ferrière Fault

La Ferrière Fault is a N-S sinistral strike-slip fault extending from the village of La Ferrière to the Val-de-Ruz (Figure 4.3). Seismic activity linked to this fault is located near the Col de la Vue des Alpes. Except for 1 event (24.01.2021), events belong to Cluster 5, beginning its activity in July 2020 and lasting until September 2020, the source origin is reactivated once on the 30th of June 2021 (Figure 4.7). The average depth of events in Cluster 5 is around 3.5 km. The isolated event occurring in January 2021 is located north of the Col de la Vue des Alpes and is probably associated to a deeper source origin (11 ± 3.7 km depth).

While the majority of events in the area of the Ferrière Fault occur during low water conditions in summer 2020, events in January and June 2021 could be associated to high water conditions and are considered for the analytical modelling.

Saint-Blaise

A thrust fault is observed north of the village of Saint-Blaise, striking SW-NE, on which a N-S striking vertical fault cross-cutting the Chaumont anticlines terminates (Figure 4.3). Over the

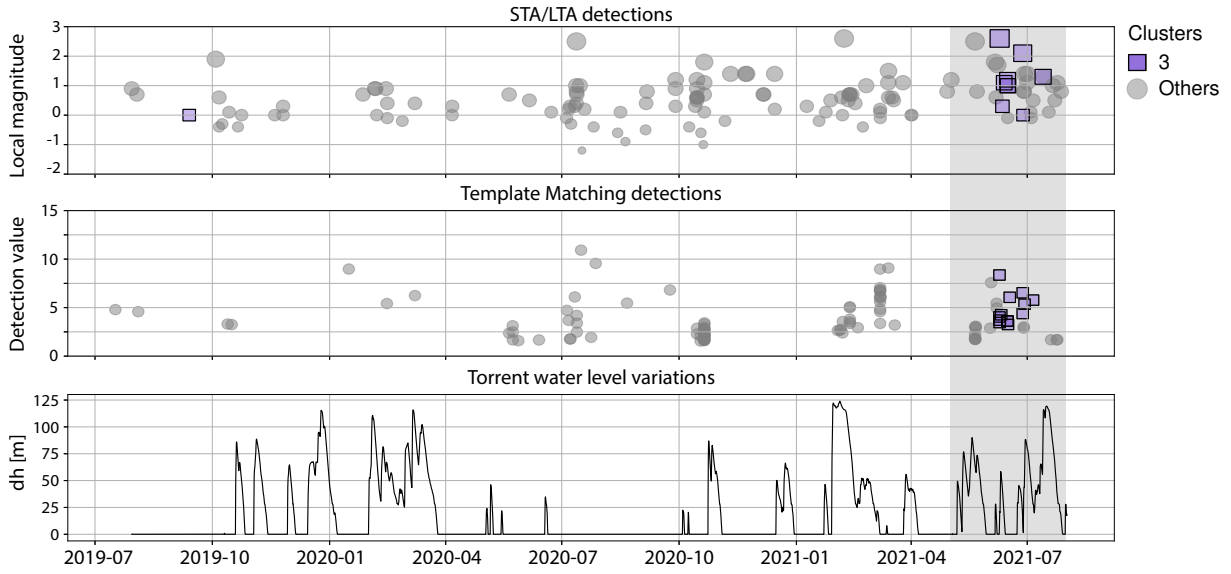


Figure 4.6: Time distribution of seismicity over July 2019 and July 2021. Black contoured squares correspond to events occurring near the Neuchâtel and Monruz Fault and discussed in the text. Color corresponds to the cluster. Shaded area refers to the period considered for the analytical modelling.

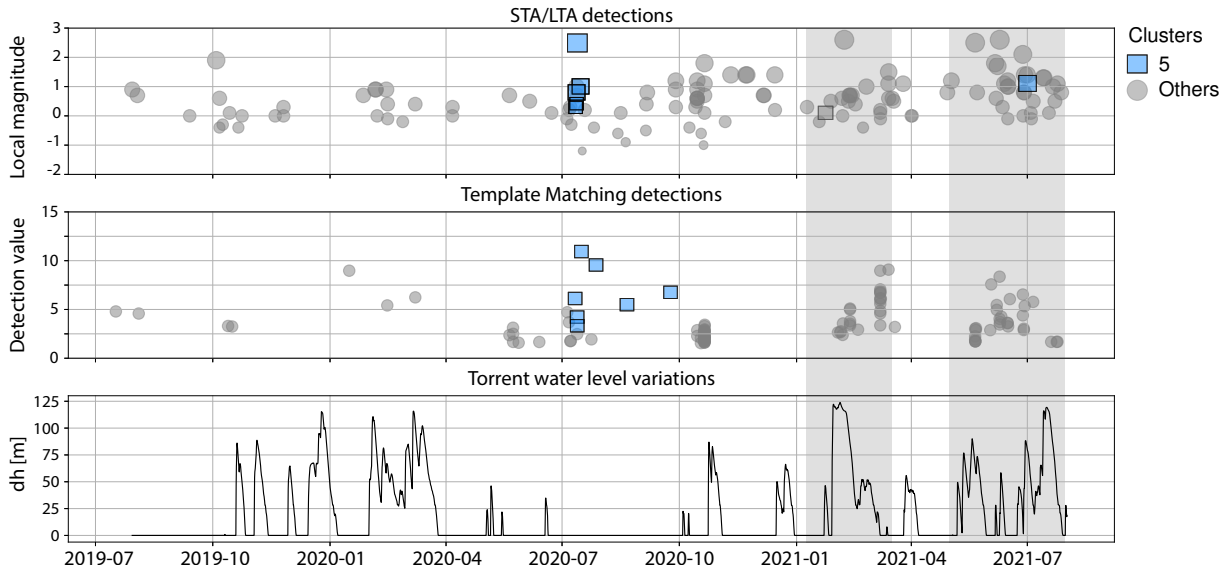


Figure 4.7: Time distribution of seismicity over July 2019 and July 2021. Black contoured squares correspond to events occurring near the Ferrière Fault and discussed in the text. Color corresponds to the cluster. Shaded areas refer to periods considered for the analytical modelling.

2 years of the monitoring period, seismic activity in this area only occurred once in February 2021. Events have an average depth of 2km and, except for 1 event (16.02.2021), they all belong to Cluster 9. These events occurred very rapidly after the maximum water level variation in the Torrent sinkhole, during very high water conditions (Figure 4.8).

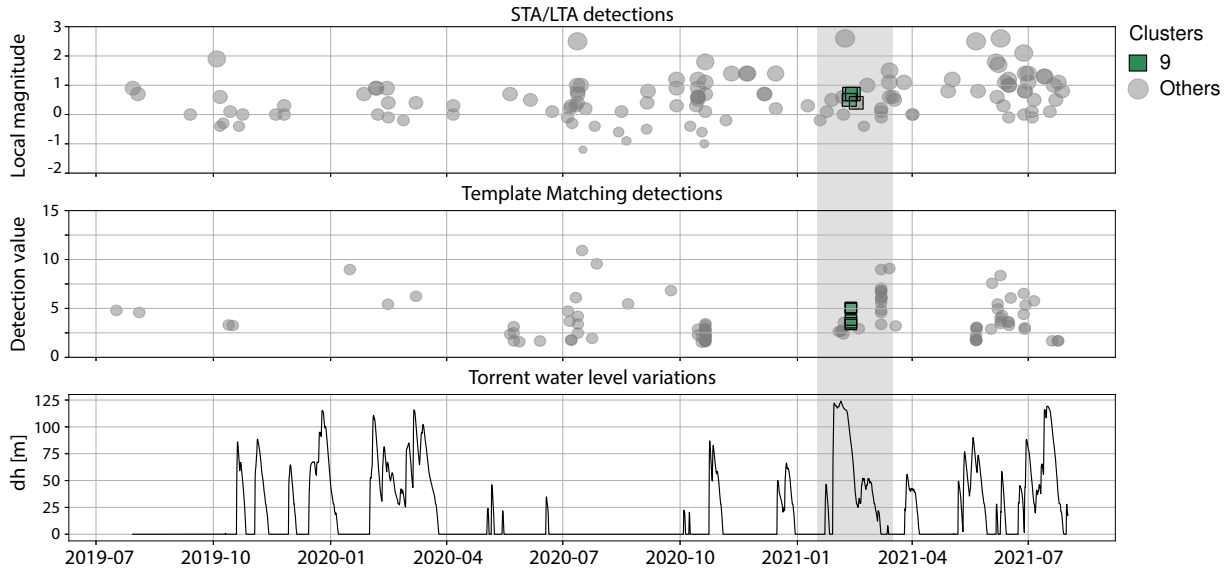


Figure 4.8: Time distribution of seismicity over July 2019 and July 2021. Black contoured squares correspond to events occurring near the village of Saint-Blaise and discussed in the text. Color corresponds to the cluster. Shaded area refers to the period considered for the analytical modelling.

La Neuveville

Near La Neuveville village, a backthrust fault is identified (Schori et al., 2015), and to the NW of the village, lies a possible N-S striking vertical fault (Figure 4.3). Seismic activity is recorded in February, March and July 2021. In February and July, events belong to Cluster 17, the event in March is considered to have another source origin as it does not belong to Cluster 17. Depth of these events is around 2-2.5 km but the error is important ($\pm > 10$ km). Except for the isolated event in March 2021, events occur under very high water conditions and are thus considered as rain-triggered (Figure 4.9).

Analytical model

We consider a semi-infinite environment with a hypocenter at depth z_{foc} below sea level and a varying hydraulic head Δh within the karstic massif (Figure 4.10). The pressure P induced by the increasing groundwater level is calculated using the groundwater level variations Δh from the Torrent sinkhole and Eq. (4.1). This pressure corresponds to the pore pressure increase at focal depth for the case of a direct connection. For the diffusion process, pressure P starts to diffuse from the bottom of the karstic network towards focal depth, over a distance L in a medium with hydraulic diffusivity D . To accommodate for the error in focal depth as well as the uncertainty regarding the altitude of the bottom of the karstic network, we apply our model to a set of distances. We also test different values of hydraulic diffusivity. The pore pressure variation at a distance L from the bottom of the karstic network at time t is calculated using the Eq. (4.7), with a step length dt of 6 hours. For the diffusion process, the history of pressure increase and decrease is important, thus we duplicate the two years of groundwater level measurements 5 times in order to have the equivalent of 10 years of fluctuating groundwater levels mimicking

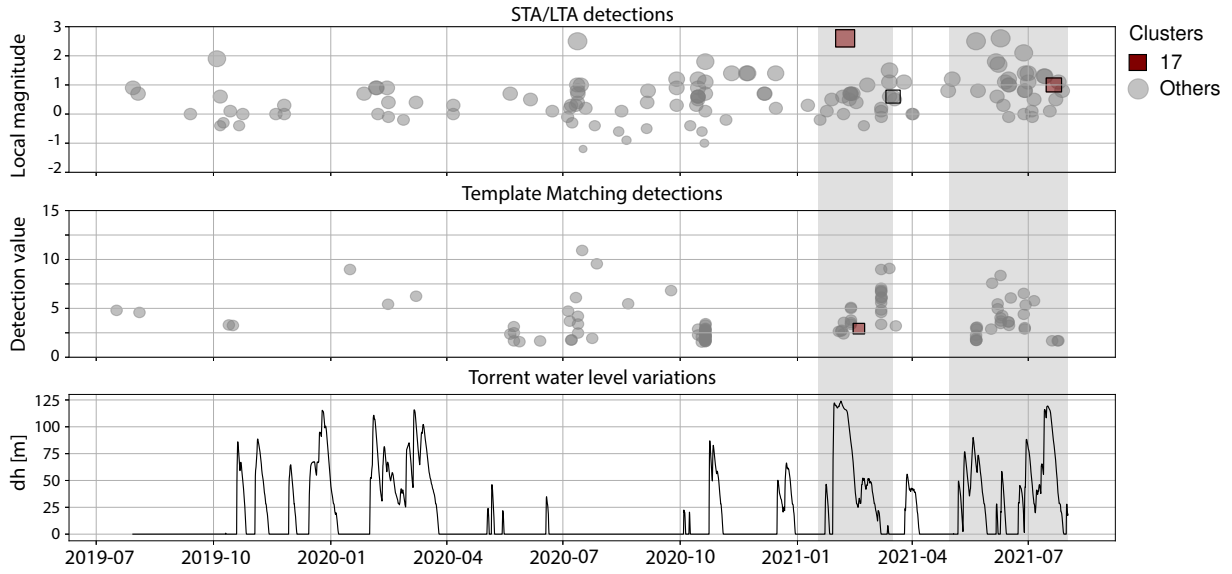


Figure 4.9: Time distribution of seismicity over July 2019 and July 2021. Black contoured squares correspond to events occurring near the village of La Neuveville and discussed in the text. Color corresponds to the cluster. Shaded area refers to the period considered for the analytical modelling.

Table 4.1: Values for the analytical modelling

Parameters	Values
Skempton coefficient B	0.2, 0.4, 0.6, 0.8
Poisson's ratio ν	0.3
Distance L [km]	0.1, 1, 2, 3.5, 5
Hydraulic diffusivity D [m/s ²]	0.1, 1, 10, 100
Water density ρ [kg/m ³]	997
Loaded radius R [km]	1

initial quasi steady state conditions in agreement with long term water fluctuation at the Torrent sinkhole. Values for the analytical model are presented in Tab. [4.1](#)

In a first step, we are interested in the pore pressure variation resulting from the groundwater level fluctuation within the Torrent sinkhole for floods during which seismicity is detected and occurs in the vicinity of known deformation structures. The evolution of the pore pressure at focal depth depends on the mechanism that is considered, either through direct hydrogeological connection between the karstic network and focal depth, through poro-elastic deformation or through the diffusion process. In a second step, we evaluate which mechanism is more likely to be responsible for the rupture of the different earthquakes. If the earthquake occurs during a period with a groundwater level fluctuation ≥ 90 m and that the level is still rising, then the mechanisms producing a direct pore pressure change can be considered. If the earthquake occurs within 48 hours following a hydraulic head ≥ 90 m, then the diffusion process is more likely to be responsible for the rupture. To evaluate which diffusion scenario fits best the observation, we assume that rupture must occur at time of a rising pore pressure or during a punctual maximum, all other scenarios are assumed invalid. Moreover we consider either isolated earthquakes or the

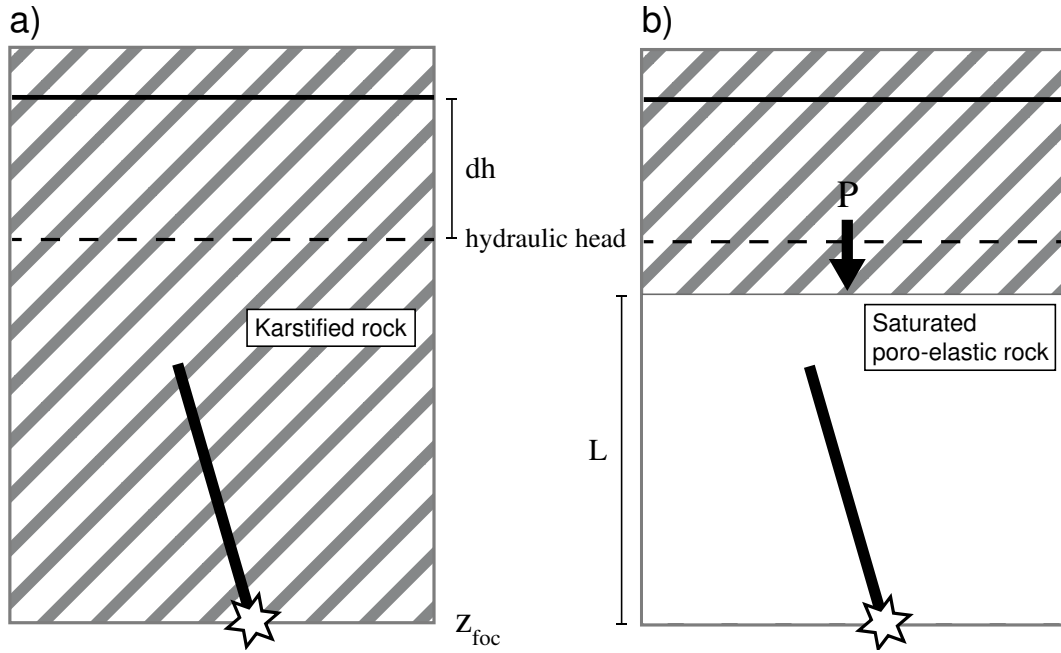


Figure 4.10: Schematic model representing a) an extended karstic network between the surface and the focal depth (direct connection) and b) a karstic network overlying a saturated poro-elastic medium without direct hydrogeological connection with focal depth (poro-elastic deformation and diffusion process).

first earthquake of a cluster. If an earthquake sequence starts during low waters, we do not consider the following events as rain-triggered.

4.4 Results

The response of pore pressure to an increasing groundwater level in the Torrent sinkhole is presented in Figures 4.11-4.14, for 4 different periods of high to very high waters, during which seismicity was observed. Earthquakes are shown on the same panels as the input signal on Figures 4.11-4.14. Potentially rain-triggered earthquakes are represented with a black cross, while other events are shown with different symbols according to their cluster. An ID is given to earthquakes considered as rain-triggered for which further information can be found in Table 4.2. Values for every scenario of hydraulic diffusivity and distance, for every considered earthquake can be found in Appendix M. Pore pressure variation resulting from the direct connection process are of the order of the MPa for a hydraulic head increase of 100 m, while it stays below the kPa for the poro-elastic effects. In both cases the response is direct and the maximum pore pressure change occurs simultaneously with the maximum hydraulic head increase. Pore pressure changes resulting from the diffusion process ranges between a few hundreds of kPa for smaller values of diffusivity ($0.01-1 \text{ m/s}^2$) and larger distances ($\geq 1 \text{ km}$) to over 1 MPa for smaller distances ($\leq 1 \text{ km}$) and larger values of hydraulic diffusivity ($10-100 \text{ m/s}^2$). A detailed description of the timing of the earthquake occurrence with respect to groundwater conditions is given for the 4 periods of high to very high groundwater levels in order to identify which mechanism is more

Table 4.2: Earthquakes considered as rain-triggered and their location. C16-2 is marked with a star (*) indicating it is a detection made through template matching. Its location is the same as its template, event C16-1. Events EQ-Yv, C8-1 and C8-2 are not within analyzed fault zones but discussed further in the text. Depth and error are in km.

ID	Origin time	lat.	long.	z	Err. z	M_L	Zone	Flood
EQ-1	21.10.19 02:44	46.921	6.805	0	> 5	-0.4	Lessy	2019-Q4
EQ-2	05.02.20 23:07	46.849	6.825	8	> 10	0.9	La Lance	2020-Q1
EQ-Yv	06.02.20 23:20	46.805	6.719	8	> 10	0	Yverdon	2020-Q1
C13-1	07.03.20 20:07	46.864	6.709	0.3	< 2	0.4	La Lance	2020-Q1
C13-2	05.02.21 14:12	46.862	6.717	0	< 2	0.7	La Lance	2021-Q1
C17-1	07.02.21 09:37	47.081	7.09	2.8	> 10	2.6	Neuveville	2021-Q1
C9-1	10.02.21 16:05	47.019	7.004	0.6	> 5	0.5	St-Blaise	2021-Q1
C8-1	11.02.21 22:22	47.002	7.188	8.1	> 10	0.7	Kerzers	2021-Q1
C8-2	11.02.21 22:45	46.99	7.179	8	> 10	0.7	Kerzers	2021-Q1
C1-1	07.03.21 15:31	46.845	6.736	4.1	< 2	0.2	La Lance	2021-Q1
C16-1	21.05.21 01:52	46.922	6.819	0	< 2	2.4	Lessy	2021-Q2
C5-1	30.06.21 23:49	47.069	6.87	12	> 10	1.1	Ferrière	2021-Q2
EQ-3	14.07.21 01:38	46.897	6.885	2.3	> 10	1.3	Lessy	2021-Q2
C16-2*	19.07.21 17:00	46.922	6.819	0	< 2	-	Lessy	2021-Q2
C17-2	21.07.21 03:01	47.069	7.071	2.1	> 10	1	Neuveville	2021-Q2

likely to have triggered seismicity. A range of possible triggering pore pressure values are given considering only valid scenarios.

Autumn 2019 Floods (2019-Q4 Flood)

Three successive floods occurred between October and December 2019 (Figure 4.11), during which the hydraulic head increased of approximately 90 m at the maximum. During the first episode of high groundwaters, an isolated earthquake occurred in the Lessy fault area (EQ-1) 24 hours after a maximum hydraulic head increase of 94 m. With such a time delay, mechanism resulting in a direct response in the pore pressure are not considered. The time delay of 24 hours suggest that the diffusion process is more likely to be behind this rupture. Considering that rupture must occur at time of a rising pore pressure or at a punctual maximum, different scenarios can be considered depending on the hydraulic diffusivity and the distance the diffusion front has to migrate. Pore pressure variation resulting from the possible scenarios range between 0.1 and 0.45 MPa for EQ-1 (Figure 4.15).

Winter 2020 Floods (2020-Q1 Flood)

A relatively extended time period of high waters occurred in February and March 2020 (Figure 4.12) during which earthquakes belonging to cluster 6 and 13 occurred in the vicinity of the western part of the La Lance fault, as well as an isolated event located on the eastern segment of this fault, near Estavayer-le-Lac (event EQ-2). While events in cluster 6 start before the flood and are not considered as rain-triggered, event EQ-2 occurs roughly 2 days after a punctual maximum of the water level in the Torrent sinkhole. The event C13-1 occurs approximately 1.5 day after the water level in the sinkhole increased of 115 m. The time lag suggest the diffusion

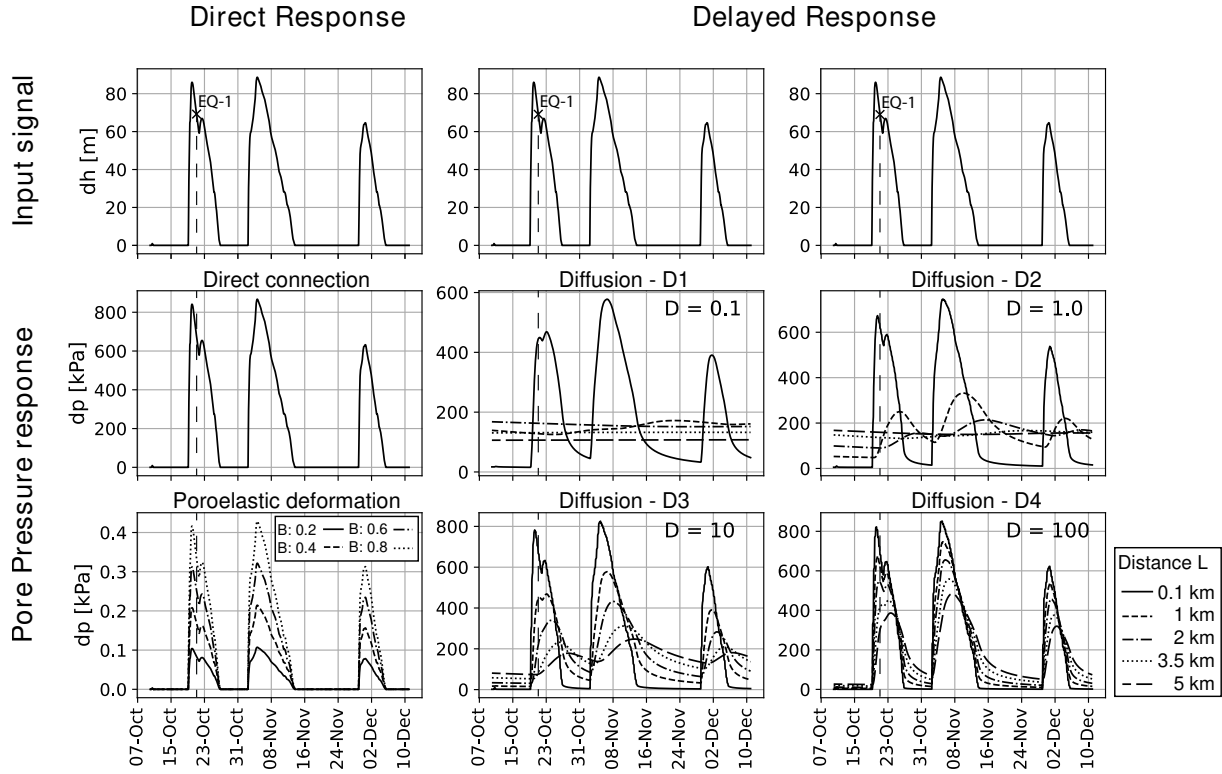


Figure 4.11: 2019-Q4 Flood. Water level variation in the Torrent sinkhole (top panels) used as pressure input signal and the corresponding response on pore pressure (bottom panels) considering the mechanism of direct connection, poroeelastic deformation (for different values of the Skempton coefficient B) and diffusion process (for different values of hydraulic diffusivity D and distances L). See text for more information.

process is responsible for an increasing pore pressure, for both events. Pore pressure increases range between 0.25 and 0.82 MPa (Figure 4.15).

Winter 2021 Floods (2021-Q1 Flood)

At the beginning of February 2021 the water level in the Torrent sinkhole increased up to 123 m and was still around 50 m at the end of the month (Figure 4.13). During that time period, several events belonging to clusters 9, 13 and 17 occurred, respectively taking place in the region of Saint-Blaise, the Lance fault and La Neuveville. The first event belongs to cluster 13 (C13-2) and occurs after 4 days of a relatively constant hydraulic head around 120 m. Shortly after the sequence of cluster 13, the first event of cluster 17 occurs on February 7th (C17-1), when the water level is slowly decreasing but still high (116 m). Three days later, the first event of cluster 9 is detected, with an associated water level of 99 m. At the beginning of March, when the water level in the sinkhole reaches its base level, cluster 1 is activated. Although events in cluster 1 do not correspond to our definition of rain-triggered earthquake, we still evaluate the pore pressure changes resulting from the different mechanisms at time of rupture because of the extended period of very high waters. For all considered events (C13-2, C17-1 and C1-1) there is a time delay between the maximum hydraulic head variation and seismicity, hence pore pressure

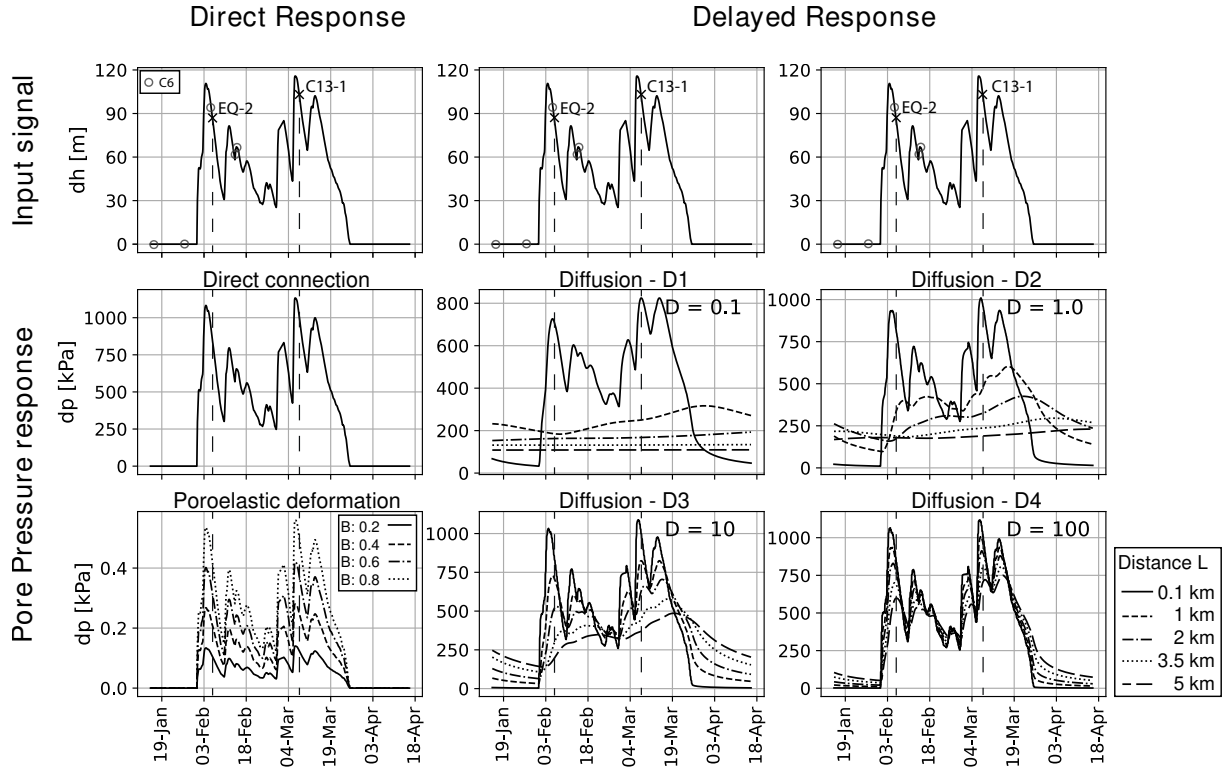


Figure 4.12: 2020-Q1 Flood. Water level variation in the Torrent sinkhole (top panels) used as pressure input signal and the corresponding response on pore pressure (bottom panels) considering the mechanism of direct connection, poroeleastic deformation (for different values of the Skempton coefficient B) and diffusion process (for different values of hydraulic diffusivity D and distances L). See text for more information.

variation most certainly results from the diffusion process. Change in pore pressure at time of rupture ranges between 0.2 and 1.1 MPa (Figure 4.15).

Summer 2021 Floods (2021-Q2 Flood)

Between May and July 2021, the hydraulic head increase reached values around 90 m in May and June and 120 m in July (Figure 4.14). After each of these high to very high groundwater levels, earthquakes were detected in the Lessy fault area (C16-1 and EQ-3), the Ferrière fault area (C5-1) and the Neuveville area (C17-1). Several earthquakes were also detected near Neuchâtel (C3) but they do not correspond to our definition of rain-triggered events and are thus not considered for further analysis. Similarly to the previous floods, earthquakes C16-1, C5-1, C16-2 and C17-2 occur within 24 to 48 hours after a peak level in the sinkhole, suggesting the rupture is induced by an increasing pore pressure resulting from the diffusion process. EQ-3, however, occurs during a punctual maximum level, when the groundwater level in the Torrent sinkhole has increased of about 120 m, suggesting that a mechanism causing a direct effect is behind this rupture. For all concerned earthquakes, pore pressure variation ranges between 0.2 and 0.75 MPa (Figure 4.15).

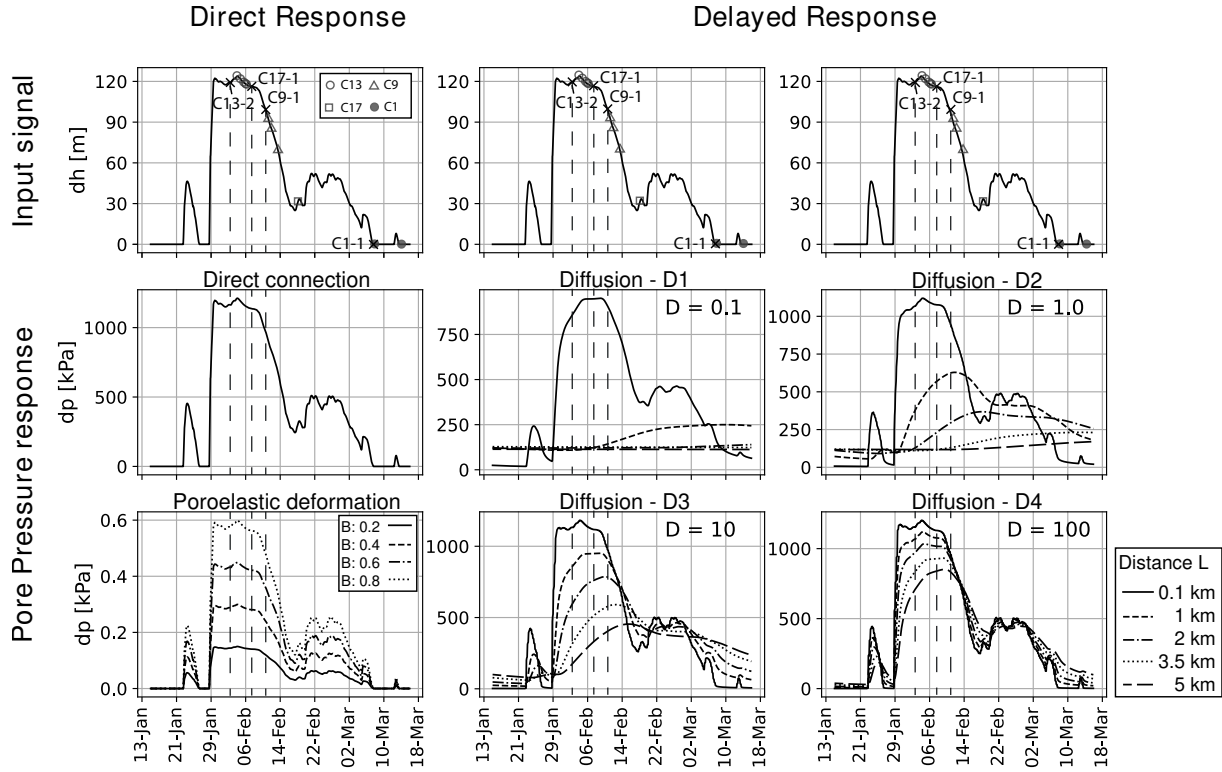


Figure 4.13: 2021-Q1 Flood. Water level variation in the Torrent sinkhole (top panels) used as pressure input signal and the corresponding response on pore pressure (bottom panels) considering the mechanism of direct connection, poroeleastic deformation (for different values of the Skempton coefficient B) and diffusion process (for different values of hydraulic diffusivity D and distances L). See text for more information.

4.5 Interpretations

The timing of the different earthquakes, occurring after a punctual maximum for the majority of the events (except for EQ-3 on the Lessy fault), seems to indicate that there is no direct hydrogeological connection between focal depth and the karstic network for the majority of the earthquakes and fault zones. Rain-induced seismicity seems to be driven mainly by the diffusion process, where the increasing pore pressure at the bottom of the karstic network (pore pressure increase from the direct connection) diffuses towards focal depth. The pore pressure variation obtained from the different possible scenarios of hydraulic diffusivity and distances all range between 0.1 and 1.1 MPa (Figure 4.15). In order to better constrain the results, we consider focal depth and its error for the different earthquakes of interest and define a pore pressure increase responsible for the rupture for each region.

La Lance Fault

La Lance fault is activated 3 times during very high water conditions with the occurrence of events in cluster 13 and EQ-2. Events in cluster 13 are located relatively close to the surface (0-300 m.b.s.l) and the error in location is lower than 2 km. EQ-2 is located further at depth around 8 km, although the error in depth is estimated to be more than 10 km. For events in cluster 13,

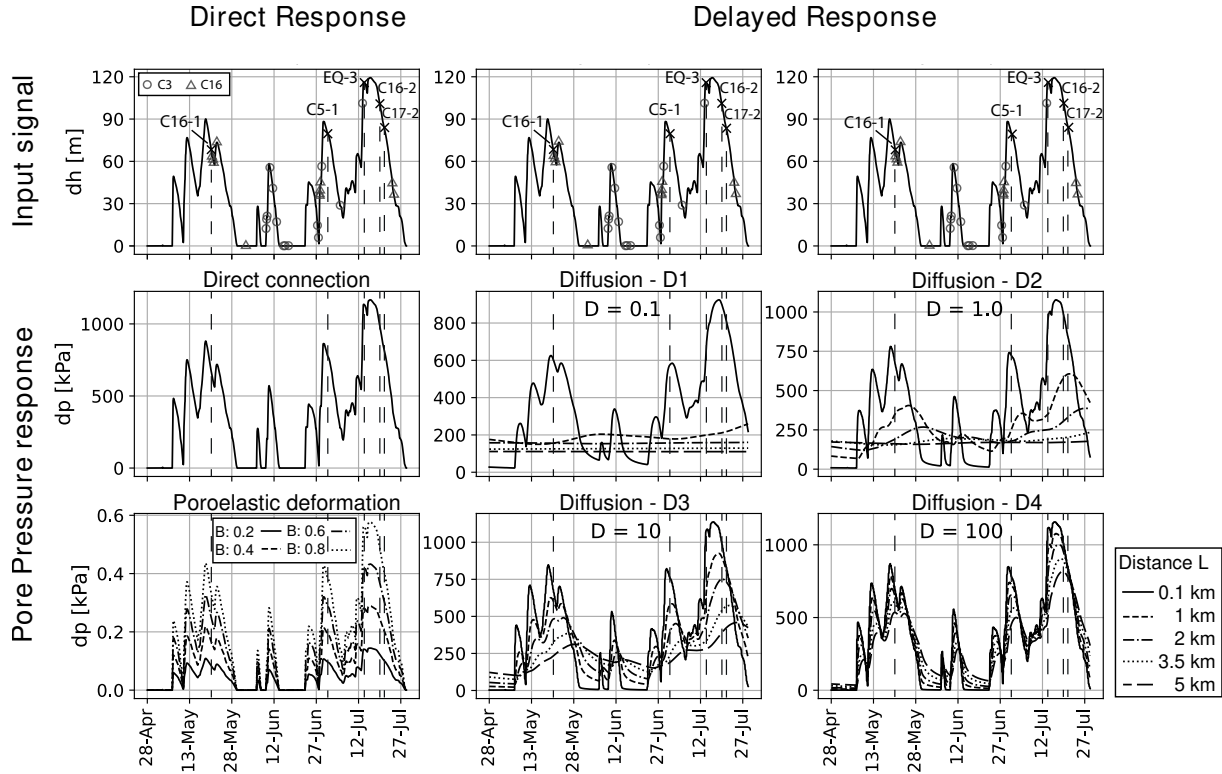


Figure 4.14: 2021-Q2 Flood. Water level variation in the Torrent sinkhole (top panels) used as pressure input signal and the corresponding response on pore pressure (bottom panels) considering the mechanism of direct connection, poroelastic deformation (for different values of the Skempton coefficient B) and diffusion process (for different values of hydraulic diffusivity D and distances L). See text for more information.

we suppose a relatively small distance for the diffusion to migrate to reach focal depth. Among the possible scenarios of a diffusing front migrating over distance between 0.1 and 2 km, only a few are valid for both events. Values range roughly between 0.2 and 0.9 MPa. The value 0.8 MPa approaches values found in scenarios of low diffusivity and short distances ($D=0.1 \text{ m}^2/\text{s}$ and $L=0.1 \text{ km}$) or higher diffusivity over larger distance ($D=10 \text{ m}^2/\text{s}$ and $L=2 \text{ km}$), which are plausible scenarios according to focal depth and its error, and allows for uncertainty regarding the extent of karstic network towards depth. For EQ-2, focal depth is not well constrained but if we consider greater distances ($L \geq 3.5$), the most important pressure changes occur in scenarios with a diffusivity of $100 \text{ m}^2/\text{s}$, with values around 0.6 MPa. For cluster 1 (2021-Q1 Flood), only scenarios with $D = 0.1 \text{ m}^2/\text{s}$, $L = 1 \text{ km}$, and $D = 1.0 \text{ m}^2/\text{s}$, $L = 3.5 \text{ km}$ could be considered, with a pore pressure change around 0.25 MPa, but since the amplitude of change is relatively small, and that these events do not fit our definition of rain-triggered earthquakes, we decide to leave them outside the analysis.

Lessy Fault

The Lessy fault shows an isolated earthquake during the 2019-Q4 Flood (EQ-1), located between 0 and 5 km depth. For small diffusivity values of 0.1 and $1 \text{ m}^2/\text{s}$, the distance the front has to migrate must be small (0.1-1 km), and pore pressure changes are of 0.44 and 0.1 MPa,

respectively. For diffusivity values of 10 and 100 m²/s, the distances can be between 1 and 3.5 km with pore pressure changes between 0.44 and 0.1, and between 3.5 and 5 km with pore pressure changes of 0.42 and 0.3 MPa, respectively. We consider a pore pressure change around 0.4 MPa as potentially sufficient to trigger this earthquake, as this value is found under different conditions, and allows for uncertainty on diffusivity and focal depth and the extent of the karstic network. During the 2021-Q2 Flood, the Lessy fault was activated twice with the occurrence of events belonging to cluster 16. These events are located at very shallow depth with an error <2 km. Scenarios of a diffusion front migrating over a small distance (0.1 km) are not valid for any diffusivity value. Scenarios with a hydraulic diffusivity of 100 m²/s is not possible for C16-2 and is hence rejected for C16-1. For distance of 1 and 2 km, pore pressure variation range between 0.15 and 0.75 MPa for diffusivities of 0.1 and 10 m²/s, respectively. However the amplitude of change in pore pressure for a diffusivity of 0.1 m²/s is very limited, suggesting that it is probably not a valid scenario. The scenario that shows the less differences of pore pressure variation between C16-1 and C16-2, is a distance of 1-2 km with a diffusivity of 1 m²/s, giving values between approximately 0.2 and 0.6 MPa, with an average value around 0.35 MPa as a triggering pore pressure for events in cluster 16. Although this result is not well constrained, it is around the same value as the one defined for EQ-1, suggesting that the Lessy faults triggering pore pressure threshold is around 0.3 to 0.4 MPa. For EQ-3, located further to the southeast in comparison to EQ-1, C16-1 and 2, the pore pressure variation can result either from a direct hydrogeologic connection between the karstic network and focal depth, or through a diffusion process considering larger migrating distances but a high hydraulic diffusivity. Considering all possible scenarios, pore pressure variation ranges between 0.4 and 1.1. If we consider EQ-1, C16-1 and C16-2 and a triggering pore pressure around 0.3-0.4 MPa, then a valid scenario for EQ-3 would be a diffusion front migrating over a distance of 3.5 km within a medium with diffusivity of at least 10 m²/s, and even more for a distance around 5 km.

La Ferrière fault

The Ferrière fault ruptured once during the 2021-Q2 Flood with an event belonging to cluster 5. Events in cluster 5 have been relocated to a depth around 3.5 km. Only scenarios with diffusivity of 10 and 100 m²/s are possible for such distances ($L > 2$ km), and values are between 0.2 MPa and 0.5 MPa, respectively. The value of 0.4 MPa is selected as the threshold pressure to trigger event C5-1 as it represents either shallower distances for the diffusion front to migrate (around 2 km) in a medium with a diffusivity of 10 m²/s, or greater distances (around 5 km) in a medium with higher hydraulic diffusivity (100 m²/s). This value allows for uncertainty regarding the depth, the extent of the karstic network at depth and the hydraulic diffusivity.

Saint-Blaise

The Saint-Blaise area showed a succession of seismic events at the beginning of February 2021, all belonging to cluster 9 and occurring after an extended period of very high waters (+120 m) (2021-Q1 Flood). C9-1 is located at a depth of 600 m.b.s.l., but with a relatively high error (>5 km). Other events in cluster 9 are located around 2 km. Although the depth is not well constrained, scenarios can be excluded considering the first event (C9-1) must occur during a

time of increasing pore pressure or a punctual maximum. Hence scenarios with a hydraulic diffusivity of $0.1 \text{ m}^2/\text{s}$ and $100 \text{ m}^2/\text{s}$ are not considered. For a hydraulic diffusivity of $1 \text{ m}^2/\text{s}$, values of 0.6 and 0.3 MPa are found, respectively for distances of 1 and 2 km, for a hydraulic diffusivity of $10 \text{ m}^2/\text{s}$ values range between 0.75 and 0.38 for distances between 2 and 5 km. An average value around 0.4 MPa is considered as the triggering pore pressure increase for the Saint-Blaise area, as it represents either distances between 1-2 km for a diffusivity around $1 \text{ m}^2/\text{s}$ or greater distances ($\geq 5 \text{ km}$) for a diffusivity around $10 \text{ m}^2/\text{s}$, and thus allows for uncertainty regarding hydraulic diffusivity, focal depth and the extent of karstic network towards depth.

La Neuveville

During the flood of February 2021 (2021-Q1 Flood), the first event of cluster 17 (C17-1) occurs when the water level is +117 m above the base water level in the sinkhole, after a maximum of +124 m, 3 days earlier. Although the water level is decreasing at time of rupture and that the 48 hours limit is exceeded, C17-1 is still considered as potentially rain-triggered because of the extended time period of very high waters (13 days). Thus, we reject the direct effect of direct hydrogeologic connection and poro-elastic deformation but consider the possible delayed effect from the diffusion process and the impact of such an important pressure step. Events in cluster 17 are located around 2.5 km depth but with an important error ($>10 \text{ km}$), thus we consider any possible distance. With a hydraulic diffusivity of $0.1 \text{ m}^2/\text{s}$ the pore pressure increases of 0.95 MPa for a distance of 0.1 km between the bottom of the karstic network and focal depth, for greater distances the diffusion front does not reach focal depth on time. For a hydraulic diffusivity of $1 \text{ m}^2/\text{s}$, the pore pressure increases of 0.55 and 0.23 MPa for distances of 1 and 2 km, respectively. For a hydraulic diffusivity of $10 \text{ m}^2/\text{s}$ at distances of 1, 2, 3.5 and 5 km, the pore pressure increase is of 0.95, 0.76, 0.51 and 0.33 MPa, respectively. For a hydraulic diffusivity of $100 \text{ m}^2/\text{s}$ and distances of 3.5 and 5 km, the increase in pore pressure is of 0.92 and 0.84 MPa, respectively. The second events C17-2 occurs during the 2021-Q2 Flood. Results show smaller values for the same scenarios. A pore pressure increase around 0.5 MPa could represent either a distance between 1-2 km and a hydraulic diffusivity of $1 \text{ m}^2/\text{s}$ or greater distances, between 3.5 and 5 km, with a hydraulic diffusivity of $10 \text{ m}^2/\text{s}$. This value would also be reached in the case the hydraulic diffusivity is $0.1 \text{ m}^2/\text{s}$ and the distance between 0.1 and 1 km. Thus, 0.5 MPa is considered as the triggering pore pressure for the Neuveville area.

4.6 Discussion

Figure [4.15](#) summarizes the considered pore pressures responsible for triggering the events identified as rain-triggered. Triggering pore pressure values range from 0.3 and 0.8 MPa. Values are between 0.3 and 0.5 MPa for the Lessy and Ferrière Fault as well as for the Saint-Blaise and Neuveville area, and are slightly higher for the La Lance Fault with value between 0.6 and 0.8 MPa. These values are derived using information about focal depth, in an attempt to better constrain possible scenarios. Although considering focal depth slightly reduces the range of possible pore pressure values, a large uncertainty remains regarding the extent of the karstic network and hydraulic properties of the crust. Thus, the preferential pore pressure value

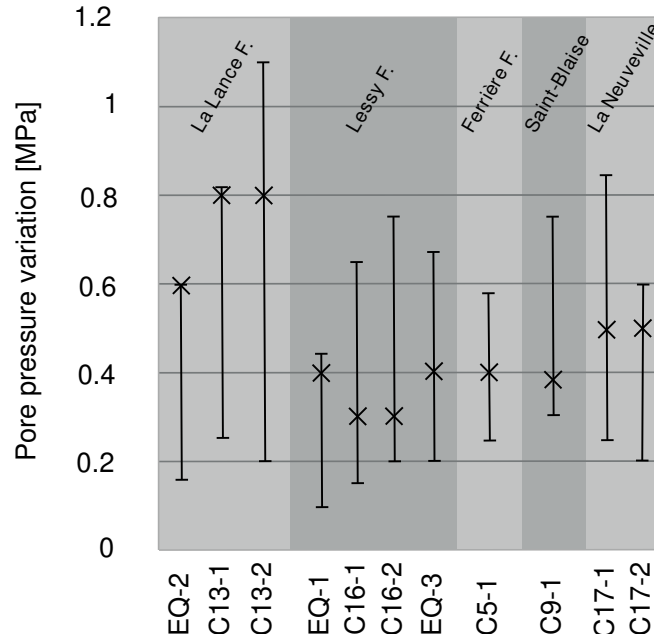


Figure 4.15: Plot showing the range of possible pore pressure variations considering all possible scenarios of diffusivity and distance for the considered earthquakes. Bars show the minimum and maximum values and the cross shows the preferred value.

mentioned for each earthquake, represents rather a possible response of pore pressure to an increasing groundwater level, without specifying any distance or diffusivity. Focal depth has been used to better constrain the distance the diffusion front has to migrate, while the lateral distribution of the epicenters around the recharge area is not mentioned. For the majority of events, their epicenters are distributed relatively closely to the recharge area justifying the use of the focal depth as only indicator for a distance scenario. However, 2 events, EQ-2 and EQ-3 discussed above, as well as the second event of C9 (hereafter C9-2), events in cluster 8 and one isolated east of Yverdon-les-Bains (not mentioned earlier because located outside the analyzed fault zones, hereafter mentioned as C8 and EQ-Yv), occur further away, at a lateral distance from the karstic recharge area of a few kilometers, between 4 and 9 km, for the events EQ-2, EQ-3, C9-2 and EQ-Yv and up to 14 km for C8, suggesting that the lateral distance between the recharge area and a potentially rain-triggered event can be much greater than depth, and that it must be considered when discussing different scenarios of diffusivity and distances.

The following section is divided in three subsections. At first, we discuss how the pore pressure increase at rupture can be used to evaluate fault criticality. Secondly, we discuss why seismicity is not as regular as expected, knowing that the defined triggering pore pressures are reached several times over the monitoring period without initiating a rupture every time. Finally, we discuss general observations regarding seismicity with respect to the geological and hydrogeological context of the study area.

Fault criticality

For the La Lance fault, higher triggering pore pressure values suggest that the fault is less sensitive to groundwater level fluctuations, in comparison to the Lessy Fault, which appears to

be sensitive to a pore pressure increase as low as 0.3 MPa. Similarly, the Ferrière Fault together with the Saint-Blaise and Neuveville area show a higher sensitivity with values around 0.4 MPa. Fault sensitivity to groundwater fluctuation can be seen as an indicator of its criticality: the more it is sensitive, the more it is critically stressed. However, since the defined triggering pore pressure is defined based on relatively few measurement for the majority of the faults (i.e. small amount of rain-triggered earthquakes), it is dubious to compare the level of criticality of the different fault zones solely with these values. A larger catalog of the study area spanning over several years would be necessary to better constrain this value, and could potentially help to identify which scenario of diffusivity and distance is taking place.

Fault criticality can also be estimated from the strike of the fault and the angle its normal makes with respect to the maximum principal stress orientation (θ between 0 and 90°), by calculating the ratio between the resolved shear and the effective normal stress on the faults (σ'_n). This is referred to as the slip tendency (Morris et al., 1996):

$$T_s = \frac{\tau}{\sigma'_n} \quad (4.8)$$

The higher the value of T_s , the higher the slip tendency. Assuming the intermediate stress is co-planar with the fault plane, then:

$$\sigma'_n = \frac{\sigma'_1 + \sigma'_3}{2} + \frac{\sigma_1 - \sigma_3}{2} \cos(2\theta) \quad (4.9)$$

$$\tau = \frac{\sigma_1 - \sigma_3}{2} \sin(2\theta) \quad (4.10)$$

where σ'_1 and σ'_3 are the effective stresses. Without information on the stress magnitudes, we can estimate the slip tendency by setting a stress ratio $R = \sigma'_3/\sigma'_1$, thus $\sigma'_3 = R\sigma'_1$. Substituting this in Eqs (4.9) and (4.10) and isolating the term $\sigma'_1/2$ gives, after simplification:

$$T_s = \frac{(1 - R) \sin(2\theta)}{(1 + R) + (1 - R) \cos(2\theta)} \quad (4.11)$$

In our case, considering faults are sub-vertical and thus co-planar with the intermediate principal stress, the maximum principal stress is horizontal. Fixing the orientation of $\sigma_1 = 135^\circ\text{N}$, which is coherent with regional observations (Rabin et al., 2018), and knowing the strike of the faults (between 0 and 180°N), we can calculate the slip tendency using Eq. (4.11) for a given R . Theoretically, faults with the highest slip tendency would be the most sensitive to pore pressure variations and vice versa. Figure 4.16a shows the normalized slip tendency for 3 different values of R as a function of the strike, as well as the theoretical pore pressure needed to initiate a slip, considering the maximum and minimum values of pore pressure variations found earlier (0.8 and 0.3, respectively). We observe that the La Lance fault, which has the highest slip tendency no matter the value of R , has also the highest pore pressure variation necessary to initiate a

slip, although a much smaller value would be expected. For all other faults, the observed pore pressure increase at rupture are more or less coherent with the expected values, within a range of 0.1 MPa. Important to note however, is that the fault considered for La Neuveville is most likely not sub-vertical as attested by the work of Schori et al. (2015) and that this method is inappropriate for this fault.

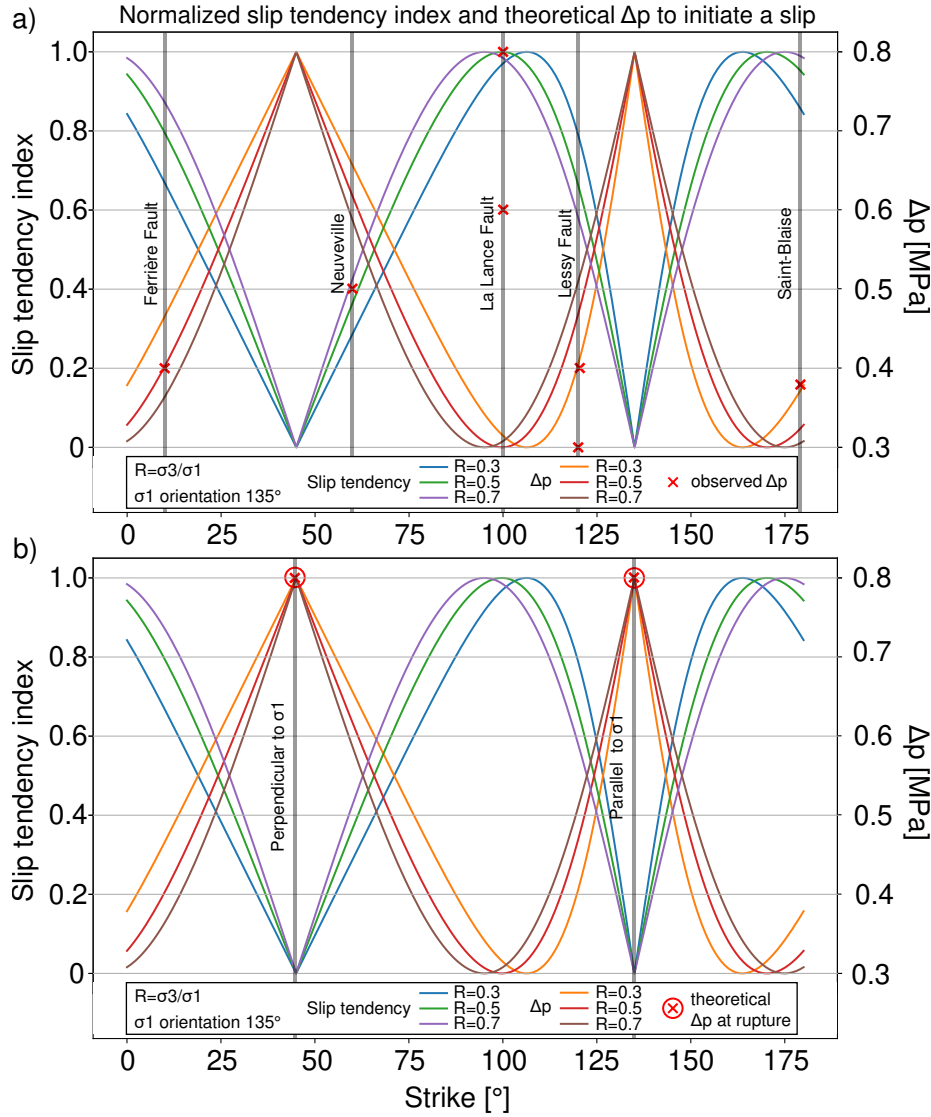


Figure 4.16: Normalized slip tendency as a function of the strike for different values of R (blue, purple and green lines), 1: very likely to slip, 0: very unlikely to slip. Theoretical pore pressure necessary to initiate a slip for a given fault orientation, with minimum and maximum based on the observed values (red, orange and brown lines). When the slip tendency is high, the theoretical pore pressure increase necessary to trigger a slip is low. When the slip tendency is low, the theoretical pore pressure increase necessary to trigger a slip is high. a) Vertical grey lines show the strike of the considered faults and red crosses are the observed pore pressures at rupture (Δp found from the analytical modelling). b) Vertical grey lines represent the strike of fault segments either perpendicular or parallel to the orientation of σ_1 , the circled red cross highlights the theoretical pore pressure variation necessary to trigger a slip on such faults.

The discrepancies between the observed triggering pore pressure at the La Lance fault in comparison to the theoretical pore pressure values could either lie in the original supposition that the maximum principal stress has the same orientation over the entire study area or in the assumption that the slip is occurring on the major fault segment oriented 100°N. In both cases, changing the orientation of the strike of the fault or the orientation of the maximum principal stress modifies the slip tendency. Figure 4.16b illustrates what would be the theoretical pore pressure for a fault segment oriented either perpendicular or parallel to the maximum principal stress, with a strike of 45°N or 135°N, respectively. The situation would be the same if the maximum principal stress was oriented 100°N (parallel to La Lance) or 190°N (perpendicular to La Lance). Thus, either local stress heterogeneities due to, for example, topography or resulting from heterogeneities in the fault zone, or the orientation of the fault segment with respect to the orientation of the maximum principal stress could explain the high values of pore pressure necessary to trigger a slip on an a priori highly critical fault.

Rain-triggered events

Values below 0.5 MPa are reached several times over the monitoring period, in many different scenarios of diffusivity and distances. Higher values (>0.8 MPa) are also found more than once, especially for scenarios of smaller distances and hydraulic diffusivity greater than $1 \text{ m}^2/\text{s}$, yet seismicity is not detected every time these values are reached. A possible reason could lie in the fact that all earthquakes are not detected. It is estimated that the magnitude of completeness (M_c) of the seismic catalog is between 0.5 and 1, which means that below these values, the catalog is incomplete. Considering the most conservative value of 1 and the magnitudes of events identified as rain-triggered (Table 4.2), almost two-thirds of the events have magnitudes below M_c , suggesting that rain-triggered events in the area are of low magnitude, and that all events are not detected. Additional reasons of why seismicity is not as regular as expected are discussed below.

Input signal

The water level in the Torrent sinkhole is used for the identification of rain-triggered earthquakes, as well as for calculating the pore pressure variations. These measures are a local indication of the hydraulic gradient, and local groundwater level fluctuations of higher or lower importance cannot be excluded. Reasons for these local differences can either be due to locally extreme meteorological phenomenon and/or to differences between the drainage capacity of the surface cover. A bare karst will drain more easily water than a silty sedimentary cover, where most of the water will runoff. A striking example of how the sedimentary cover can influence the infiltration in our study area is the flash flood that occurred in Cressier (Neuchâtel) on June 22nd 2021, where approximately 40 mm of rain fell within 30 minutes on the catchment area of a small stream. Almost half of this watershed is covered in sediments from the last glaciation comprising silty sands and gravels, hindering a rapid infiltration within the karst conduits. On June 22nd 2021, the rain began around 7:30 pm and 45 minutes later, water was flowing down roads from the village carrying large amounts of sediments, destroying cars and houses. During the same period, a similar amount of rain fell further north east, on bare limestones showing

heavy signs of karstification as attested by the numerous dolines references on the geological map (see Appendix [N](#)), however, no flooding was reported in this region. Signs of karstification in the regions of the La Lance and Lessy faults, where a large amount of dolines are reported on the geological map and visible on the digital terrain model (DTM, swisstopo), suggest that the rainfall is directly drained towards depth. Dolines are also visible near the southern part of the Ferrière fault, although to a lesser extent, and north of Neuchâtel and Monruz. For the eastern part of the study area, Saint-Blaise and La Neuveville, signs of karstification on the surface are visible on some location, but a relatively large part of the catchment area of the surrounding karstic springs are covered with sediments from the last glaciation. The variation in karst density (at least at the surface), from the west to the east, is compatible with observations of the time delay between the maximum rain intensity and the maximum discharge at the spring (see Chapter [3](#)), where it is shorter for the La Lance and Dia spring than it is for the Serrière spring and the Torrent sinkhole. Considering these possible local site effects, a misidentification of the rain-triggered event is not excluded, nonetheless, all regions present at least some signs of karstification at the surface and it is reasonable to consider that water is relatively rapidly drained towards depth.

Fault properties, hydromechanical coupling processes and earthquake cycles

As mentioned in Chapter [1](#), the hydraulic properties of a fault are subject to changes over time. The seismic activity of the faults in the study area is not solely restricted to periods of high to very high waters and changes in hydraulic properties can occur following any event. For example, changes in permeability following a *dry* event could explain the occurrence or absence of a *wet* event. A permeability increase enhances the drainage capacity of the fault zone: pore pressure decreases if the fluid is drained away, or increases in the fluid flows towards the fault zone. Similarly, a decreasing permeability due to fault healing processes ([Gratier, 2011](#)) promotes fluid overpressure within the fault zone if the fluids cannot escape, or hinders the fluid to penetrate within the fault zone. Moreover, deformation may not necessary be seismic, making it difficult to link an increasing ground water level to a deformation without the adequate detection methods (aseismic slip monitoring through GPS measurement, see for example [Ozawa et al. \(2002\)](#)). Additionally, considering aseismic slip motion, the occurrence of *wet* events could be interpreted as "side effects" of a silent slip, as it is suggested from fluid injection experiments in fractured rocks, where the timing and location of seismic events do not correspond to the migration of the pressure front ([Duboeuf et al., 2017](#)). Stress redistribution around the fault or changes in the frictional resistance on the fault plane following aseismic movements could then be responsible for the occurrence of earthquakes relatively shortly after major flood, without having a direct relation with the groundwater level variations. Considering the stick-slip model for earthquake generation, if the flood occurs rapidly after the beginning of a new cycle, i.e., after an earthquake, the stress accumulated on the fault plane might not be enough to be sensitive to an increasing pore pressure, while it might be "critical" at the end of a cycle. Similarly, the stress drop during an earthquake influences the timing of the next earthquake nucleation. While the values of stress drop are not determined in this study, we use the analytical method of [Shapiro and Dinske \(2021\)](#) for the maximum stress drop estimate.

The authors assume a favourably oriented fault with no cohesion in a strike slip regime with the mean stress $((\sigma_1 - \sigma_3)/2)$, $\sigma_m \approx \sigma_v$, and where the vertical stress is insignificantly affected by the stress drop. Also assuming a static friction coefficient of 0.6 and a maximum variation of -0.1, from the static to the dynamic friction coefficient (see Chapter 1, section 1.2), they calculated the maximum stress drop as follows (see Shapiro and Dinske (2021) for complete description of the method):

$$\Delta\sigma = 0.074(\sigma_v - p) \quad (4.12)$$

where p is the pore pressure. Assuming an average depth of 2 km with a lithostatic gradient of 26.5 MPa/km, a saturated thickness of 1600 m and hydrostatic pressure gradient, then $\Delta\sigma \leq 2.74$ MPa. This maximum stress drop assumed for the most favourably oriented fault (and additional reasonable assumptions mentioned above) is in the same order of magnitude than variation of pore pressure resulting from an increasing hydraulic head (MPa scale). A stress drop in the range of the MPa, is also coherent regarding the magnitudes of the events and the size of the fault segments (Figure 4.17).

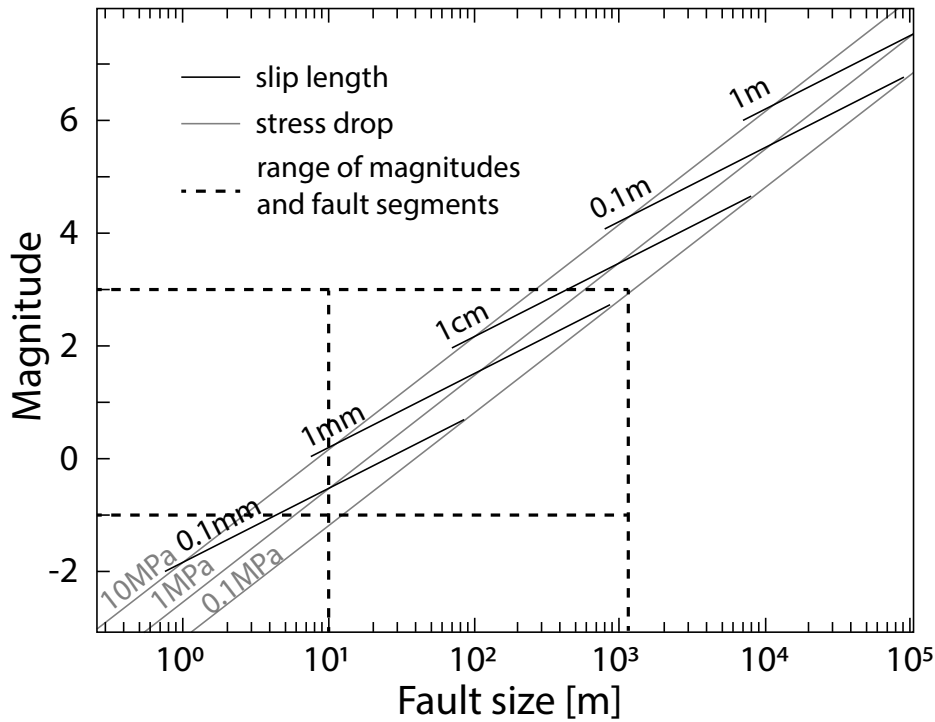


Figure 4.17: Relationship between magnitude, stress drop and fault size. Figure modified from Zoback and Gorelick (2012).

General observations

As presented in Chapter 3, the study area is located at the foot of the Internal Jura: to the south, the sedimentary layers are dipping relatively constantly to the southeast (12-15 °) while the dip is steeper (20-30 °, up to 50°) in the northern part (Figure 4.18). From reconstructed cross section (Rime et al., 2019; Schori et al., 2015; Gruber, 2017), it is thought that the units

found on top of the first anticline (Mont Aubert, Montagne de Boudry, Chaumont) are located approximately between 0.5 and 0.8 km b.s.l. below the southern shore of lake Neuchâtel. Units on top of the anticlines are mostly composed of limestone from the upper Jurassic and are highly karstified. These units represent the major groundwater reservoir of the region, and are isolated from units of the middle Jurassic by an impervious layer. A well for drinking water drilled in this aquifer, near the village of Bevaix shows an artesian pressure fluctuating between 0.5 and 0.8 MPa (data from the exploitation company, eli10). In the following, we discuss how this context 1) influences the general distribution of seismicity and 2) how, coupled with rain-triggered seismicity, it informs on deep seated groundwater circulation.

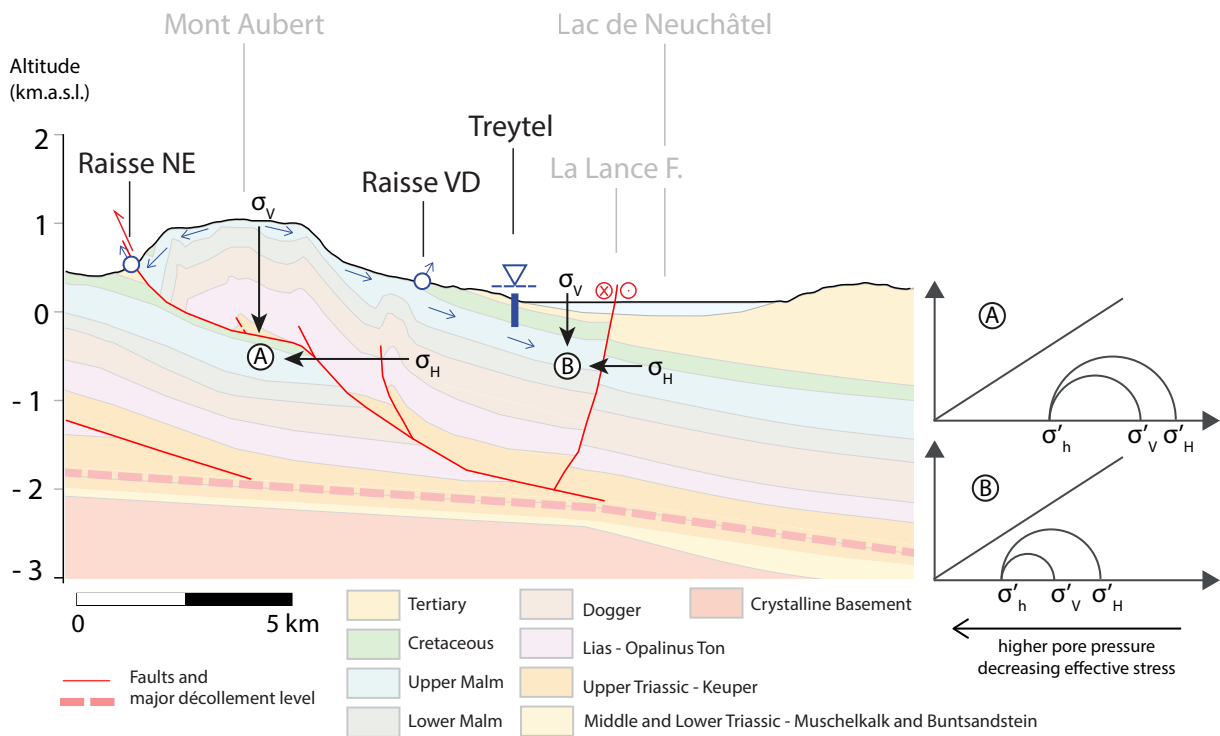


Figure 4.18: Cross section across the first anticline of the Jura Mountains and Lake Neuchâtel showing the geological structures. Water infiltrates on the mountain and flows down the slope, following the dip of the layers. Point A, below the mountain has a higher effective stress than point B, below the lake, because of a deficit in pore pressure due to the permeability heterogeneity and the presence of the thrusts (possibly acting as a barrier to fluid flow). Figure modified from Rime et al. (2019).

Seismicity distribution

From the entire catalogue, we observe that the majority of earthquakes occur below the lake, near known mapped faults outcropping on shore (Figure 4.19). Only few events occur below the mountains (clusters 3 and 5). On Figure 4.18, consider the point A, located below the anticline and the thrust fault, and the point B situated below the lake, both at the same altitude below sea level. Imagine a recharge event on the mountains, fluid pressure will propagate under the lake due to the SE dipping of the layers and permeability heterogeneity. In addition, the thrust might act as a barrier to fluid flow, favoring horizontal to sub-horizontal flow and hindering

fluids to propagate towards depth. At a same depth, rupture is more likely to occur at point B because of an excess in pore pressure (as attested by the artesian pressure at the Treytel well), resulting in lower effective stresses in comparison to point A. Thus, the stress condition at point B are more likely to generate a rupture, because closer to the failure envelop than the stress condition at point A.

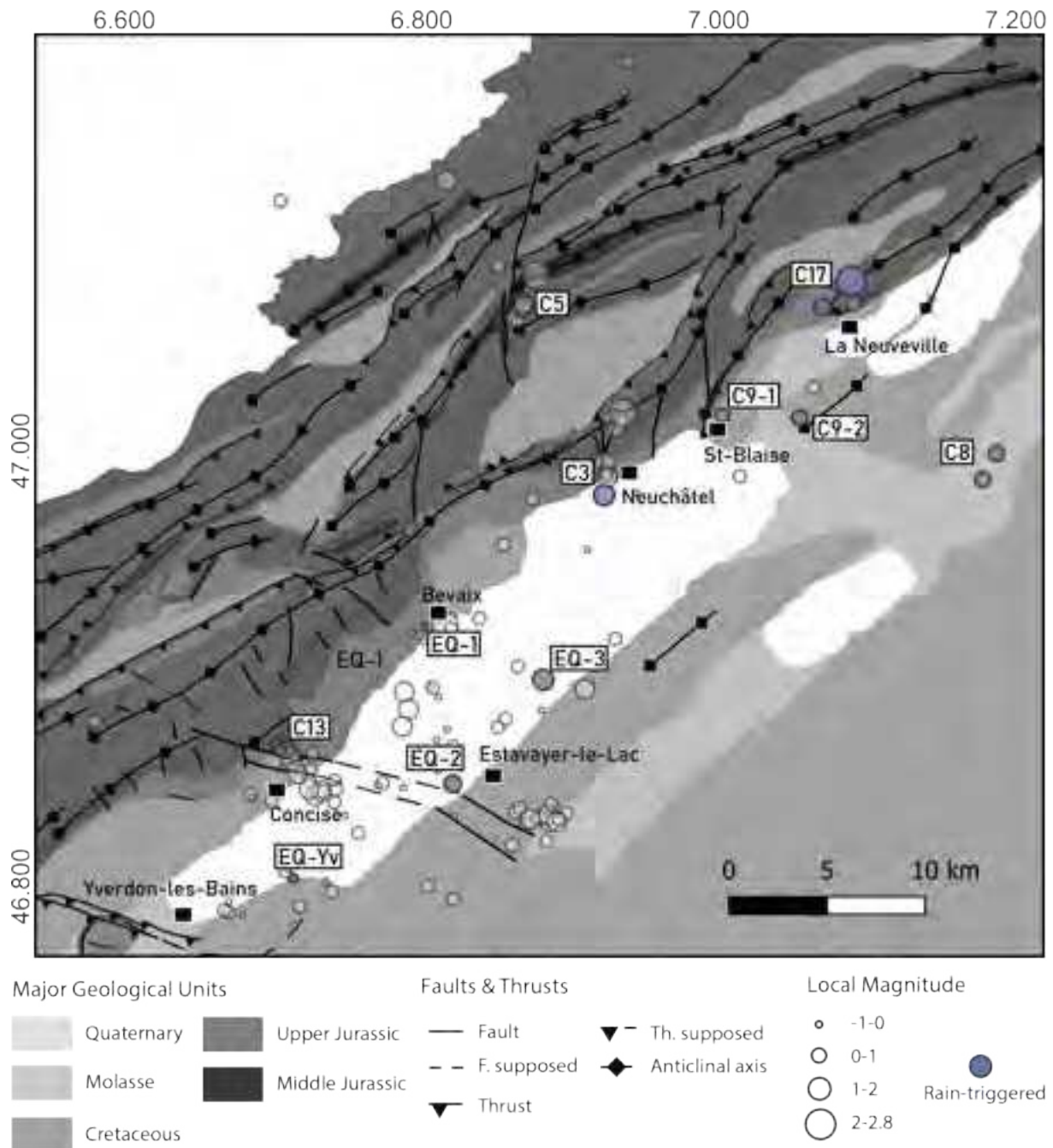


Figure 4.19: Seismicity in the study area. Highlighted in blue: rain-triggered events and their name. C3 occurred during high groundwater conditions but is part of a sequence that started during low waters, thus, it is not mentioned in the analysis. Background map compiled after the geological and tectonic map from the Federal Office of Topography, swisstopo

Groundwater circulation

Groundwater circulation depends on the hydraulic properties of the rock, especially on permeability. The stacking of aquitar and aquifer units governs the fluid flow in the sense that it directs the flow parallel to the dipping of the layers. In addition, the role of faults acting either as barrier or conduits influence the flow direction by hindering or channeling the water flow, along or across the fault. While the majority of rain-triggered events are located close to the recharge area (outcropping limestones), suggesting vertical fluid flow, a few of them are located further south, several kilometers away at depth ≥ 2 km, as it is the case for EQ-2, EQ-3 as well as event C9-2 and additional events of cluster 8 and EQ-Yv (Figure 4.19). For the analyzed events EQ-2 and EQ-3, considering the lateral distance and the depth implies a high hydraulic diffusivity to obtain an increasing pore pressure at time of rupture. Although a high hydraulic diffusivity (e.g. ≥ 10 m²/s) is hard to imagine over several kilometers depth, it can be justified in cases of high permeable pathways such as favorably oriented faults that are most likely hydraulically conductive (Barton et al., 1995) or heavily fractured and karstified rock. Assuming that fluid flow, and thus fluid pressure propagation, is concentrated in the major aquifers of the area, which are within units of the Cretaceous, the Upper Jurassic (Malm), and, to a smaller extent, the limestones of the Middle Jurassic (Dogger), and that these aquifers can be connected in highly fractured places, rain-triggered events can take place at greater depth (or greater lateral distance) in a relatively short time delay. Figure 4.20 proposes a conceptual model on how fluid pressure is propagated following the preferential paths, based on the location of rain-triggered events. Number 1) in Figure 4.20 could correspond to events C13, EQ-1 and C9-1, where pore pressure rapidly propagates at the bottom of the karstic network and then diffuses through high permeable pathways such as fault splays or jogs to reach focal depth. Number 2) could schematically represent events C5 and C17, where the same structures allow the fluids (and the fluid pressure) to reach even greater depth, eventually recharging the deeper aquifers. Number 3) could schematically represent earthquakes EQ-3 and C9-2, where groundwater flows parallel to the fault within the Malm aquifer. The presence of jogs could allow the groundwater to flow vertically and to reach the aquifer of the Dogger, and then to flow parallel to this aquifer, eventually reaching focal depth schematically represented by number 4) and corresponding to events C8, EQ-2 and EQ-Yv.

4.7 Conclusion

Combining microseismic activity together with hydrogeological data allowed us to investigate fault criticality and hydrotectonics of the Neuchâtel Jura. Using simple 1-D analytical solutions, pore pressure variation following important groundwater level increases were calculated for earthquakes occurring during high to very high groundwater conditions. Results of this analysis show that pore pressure variation range between 0.3 and 0.8 MPa. Factors influencing pore pressure variation include focal depth, the extent of the karstic network and the hydraulic properties of the crust, which are also the factors with the highest uncertainty. A higher precision in earthquake location, together with field investigation regarding the karstic network distribution could already improve knowledge about hydraulic properties of the crust, and thus

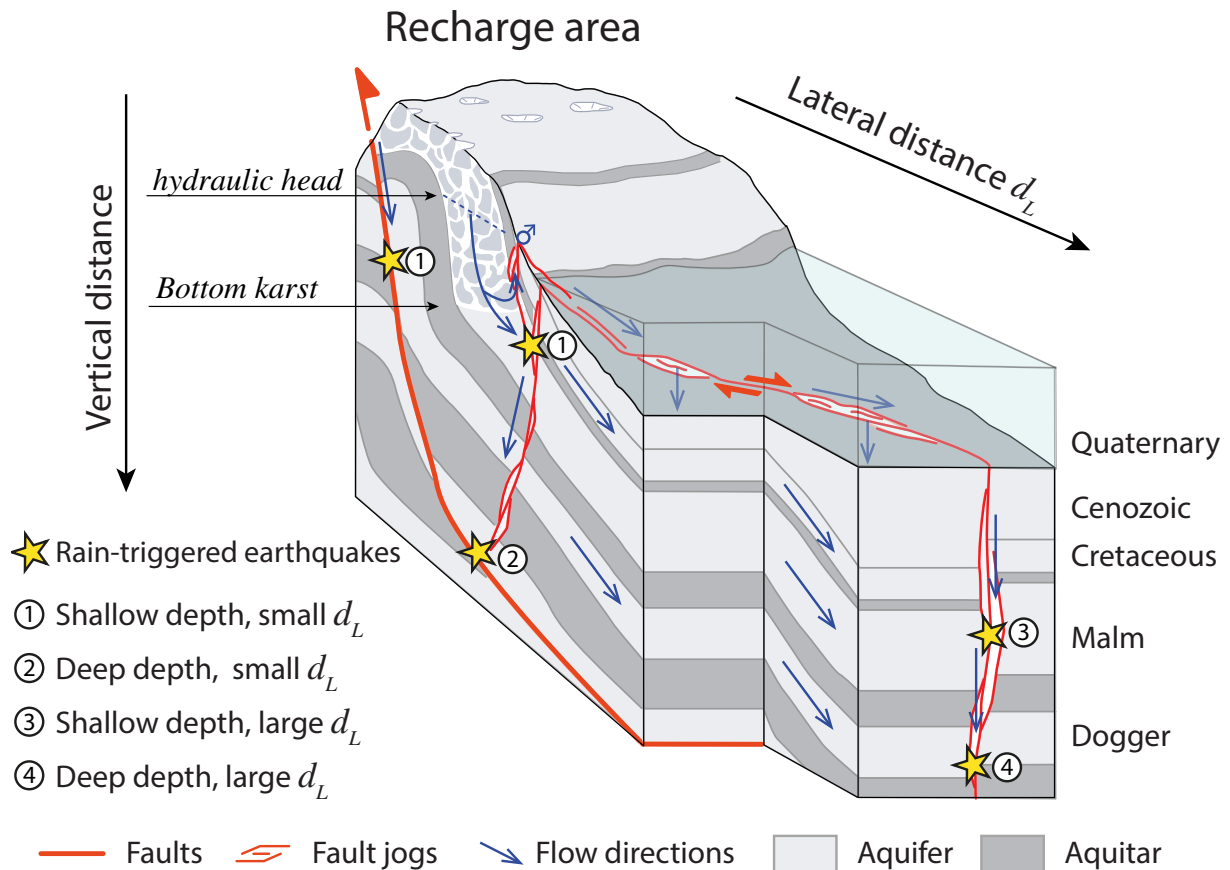


Figure 4.20: Conceptual model illustrating the groundwater circulation within the aquifers and how these can be connected in highly fractured zones leading to rapid fluid pressure propagation towards seismogenic depth. Numbers and stars refer to different scenarios of hydroseismicity. 1) Earthquake at shallow depth and short distance from the recharge area, 2) earthquake at greater depth but small distance from recharge area, where water flows along the fault, through high permeable pathways (fault splays and jogs), 3) seismicity occurs at shallow depth but large distance from recharge area, water is channeled parallel to the fault and parallel to the aquifer layers and 4) same as previous but at greater depth, where the presence of jogs might promote groundwater to flow vertically and reach the deeper aquifers.

provide a higher precision on the pore pressure variation necessary to trigger an earthquake. Nonetheless, results are within the range of values found in literature to trigger seismicity on pre-existing faults, between 0.1 and 1.1 MPa (Lockner and Beeler, 1999; Roeloffs, 1988; King et al., 1994; Saar and Manga, 2003), and are thus considered as sufficient to trigger seismicity in the Neuchâtel Jura. Specific values of pore pressure variation have been determined for different fault zones and used as an indicator for fault criticality. Interestingly, the fault zone with the highest pore pressure variation, i.e. supposedly the less critically stressed, is the fault zone, which has the highest slip tendency for a given stress state. It is also the fault zone with the highest number of earthquakes. This observation is most certainly due to a varying angle between the normal to the fault plane and σ_1 , either resulting from a varying orientation of the maximum principal stress or a misidentification in the fault strike. Evaluating fault criticality solely based on rain-triggered events seems to be insufficient for the Neuchâtel Jura, as seismicity occurs under any hydrogeological conditions and no clear correlation between high groundwaters

and the seismic rate was observed. Nonetheless, over the two years of the monitoring period, seismicity in the study area was detected 4 times out of 5 periods where the groundwater level in the Torrent sinkhole reached the 100 m threshold. This observations suggests that during very wet conditions, there is 80% chance to detect seismicity in the study area. However, the observation period should be longer to confirm this suggestion. The overall distribution of earthquakes shows that the majority of seismicity is localized at the foot of the Jura Mountains and below the lake. A deficit in the vertical load for a same hydrostatic pressure is proposed as a possible explanation for the seismic activity pattern, where smaller effective stresses are found below the lake. This hypothesis is also coherent with the local geological and hydrogeological context, where groundwater is recharged by infiltration on the mountains and flows towards the southeast, following the dip of the layers, favoring an excess in fluid pressure below the lake. Rain-triggered seismicity within this context informs on the preferential pathways for groundwater circulation, which is most certainly concentrated along the faults.

Chapter 5

Stress inversion: Present-day stress field from microseismicity and fluid pressure analysis

5.1 Introduction

Estimation of the stress field in the earth's crust is of great importance for deep underground projects, i.e. mining, oil and gas industries, geothermal exploitation, deep disposal of nuclear waste, as well as for better understanding tectonics, faulting and seismicity. Information on the stress state of a specific area helps to design and operate underground infrastructures in the most efficient way and is essential to mitigate the inherent risks of such infrastructures; i.e. induced seismicity, tunnel or borehole collapse or leakage of waste (Zang and Stephansson, 2010; Jaeger et al., 2009).

Commonly, the stresses are measured indirectly using various in-situ techniques (Zang and Stephansson, 2010) or estimated through stress inversion methods based on geological observations and geophysical data acquisitions. Most common measurement techniques based on hydraulic tests in boreholes (Haimson and Cornet, 2003; Schmitt and Haimson, 2017), on the analyses of borehole failure (Valley and Evans, 2019; Schmitt et al., 2012) or the measurement of borehole deformation while drilling (Sjöberg et al., 2003) requires access to the subsurface with boreholes, shafts or tunnels. Such methods can provide quantitative estimate of at least some components of the stress tensor at depth. However access to the subsurface is not available during the early prospecting phase of a project. Thus, at early project phase, it is suggested to derive a so-called Best Estimate Stress Model (Stephansson and Zang, 2012) based on generic data bases, tectonic information and inversion from focal mechanisms.

Stress inversion methods from focal mechanisms have been developed by several authors (Michael, 1984; Geophart and Forsyth, 1984; Angelier, 2002), and allow to determine a reduced stress tensor responsible for a given slip. The full stress tensor is defined by six independent components (e.g. three orientations and three magnitudes), and can be decomposed in isotropic and deviatoric components. The latter is the only one contributing to the shear stress since the isotropic component acts normal to any plane and is unable to produce any shearing (Zang and

[Stephansson, 2010](#)). The orientation and the sense of the shear component is not affected by any changes in the isotropic components, but by the orientation of the principal stresses and their relative ratio. Thus, instead of determining six component for the full stress tensor, only four parameters are necessary to represent the shear component (three angles and the relative principal stress ratio). The reduced stress tensor doesn't include the absolute principal stress magnitudes but contains sufficient information to compute unequivocally the resolved shear stress vector on a fault plane. Inversion method from focal mechanisms aims at identifying the best reduced stress tensor for a given set of faults, assuming all faults slipped under the same stress. Such inversion are typically based on the Wallace-Bott hypothesis (see section [5.2](#)), i.e. the assumption that the slip vector and resolved shear stress vector are co-linear. The optimal solution is defined as the one that presents the lower misfit angle between the calculated and the observed slip direction on the fault planes. In order to estimate stress magnitude, additional information must be integrated. An inversion technique based on fracture reactivation and hydraulic measurement allowing the estimation of the complete stress tensor has been proposed by [Kakurina \(2019\)](#). An adaptation to this method in order to extract absolute stress magnitude from focal mechanisms, hydrogeological observation and hydrodynamic computation is proposed in this work.

Another approach for stress estimation is to derive a so-called 1st order estimate, i.e. an estimate of the major trends of the stress state, by considering only the dominant processes controlling the stress state in the crust, namely gravity and tectonic deformations. The six components of the stress tensor, three orientations and three magnitudes can be estimated considering the following assumptions: (i) The vertical stress is a principal stress, which makes the two other principal stresses being horizontal. (ii) The maximum horizontal stress is oriented perpendicularly to specific geological structures, such as fold axis or reverse fault segments. The magnitude of the vertical stress is proportional to the weight of the overlying rock, and the magnitudes of the horizontal stresses depend on the vertical stress (gravity component) as well as on the deformation caused by tectonic forces. While the gravity component is easy to calculate, the tectonic component necessitates knowledge about the deformation and mechanical parameters. Although this method is a simplification as it considers only an elastic crust behaviour and neglects heterogeneities arising from failure, it can nevertheless include elastic heterogeneities and complex topographical effects if solved with advanced numerical methods, however this is not the purpose of this chapter. This simpler methods allows already to give a first estimation of the stress state in a given region.

This work aims at estimating the stress state in the Neuchâtel Jura using the aforementioned methods. A first estimation of the stress state is done using the simplistic approach of the gravity and the tectonic stress models, knowledge about the orientation of local geological structures and their stability helps to better constrain the possible stress scenarios. A second estimation inspired by the work of [Kakurina \(2019\)](#) is done using the inversion technique developed by [Angelier \(1979\)](#) and modified by others ([Yamaji, 2003](#)), using focal mechanisms identified in Chapter [3](#) and in literature and pore pressure variation estimates allowing for the estimation of the full stress tensor (i.e. including magnitudes). The inversion technique developed by [Michael \(1984, 1987\)](#), modified by [Vavryčuk \(2014\)](#) and implemented in the STRESSINVERSE program

(Vavryčuk, 2014) is also used and results are compared. Finally, we discuss how the results are influenced by the model's limitations and how they fit in the general tectonic context of the study area.

5.2 Methods

Gravity and tectonic stress models

Within Andersonian stress regime theory, i.e. making the assumption that one principal stress direction is vertical, three principal stresses magnitudes are used to define the stress field in the earth's crust, namely the vertical stress σ_v , the maximum horizontal stress σ_H and the minimum horizontal stress σ_h . The justification of Anderson's theory is the dominant role of gravity acting vertically on the in-situ stress state. Topographical effect can influence the stress orientation near the earth surface. Other stress sources including tectonic deformation, thermal effects and isostatic relaxation are superposing to the gravitational stresses. Considering gravitational and tectonic effects on an elastic upper crust, it is possible to derive an analytical solution to compute the stresses in the crust. From the lithostatic stress model, the vertical stress σ_v at depth z_1 reads :

$$\sigma_v = g \int_{z=0}^{z_1} \rho(z) dz \quad (5.1)$$

where ρ is the rock density, g the gravitational acceleration and z the depth recorded downward from earth surface. When the topography is not flat Eq. (5.1) is still valid at depth sufficiently large compared to the amplitude of the the topography and taking the mean elevation as the reference surface ($z = 0$). The maximum and minimum horizontal stresses also depend on the gravitational component. Assuming that there is no horizontal deformation under a given load ($\varepsilon_H = \varepsilon_h = 0$), σ_H and σ_h can be written as a function of the vertical stress (the load) and the Poisson's ratio ν (Zang and Stephansson, 2010):

$$\sigma_H = \sigma_h = \sigma_v \frac{\nu}{1 - \nu} \quad (5.2)$$

In addition to the gravitational field, Savage et al. (1992) allowed for horizontal tectonic strains ε_H and ε_h to influence the horizontal stresses and found the following relation :

$$\sigma_h = \sigma_v \frac{\nu}{1 - \nu} + \frac{E}{1 - \nu} (\varepsilon_h + \nu \varepsilon_H) \quad (5.3)$$

$$\sigma_H = \sigma_v \frac{\nu}{1 - \nu} + \frac{E}{1 - \nu} (\varepsilon_H + \nu \varepsilon_h) \quad (5.4)$$

where E is the Young's Modulus.

To avoid a super critical stress state, a maximum ratio between σ_1 and σ_3 as defined in (Jaeger et al., 2009) is set:

$$\frac{\sigma_1 - p}{\sigma_3 - p} = \left[\sqrt{\mu^2 + 1} + \mu \right]^2 \quad (5.5)$$

where $\left[\sqrt{\mu^2 + 1} + \mu\right]^2 = f(\mu)$. σ_1 and σ_3 are the total maximum and minimum stresses, respectively, p is the pore pressure and μ the friction coefficient. From Eq. (5.5), one can also calculate the maximum and minimum possible horizontal stresses, σ_{Hmax} and σ_{hmin} , respectively. In the case where $\sigma_1 = \sigma_H$ and $\sigma_3 = \sigma_v$ (thrust faulting regime):

$$\sigma_{Hmax} = f(\mu)(\sigma_v - p) + p \quad (5.6)$$

and in the case $\sigma_1 = \sigma_v$ and $\sigma_3 = \sigma_h$ (normal faulting regime):

$$\sigma_{hmin} = \frac{(\sigma_v - p)}{f(\mu)} + p \quad (5.7)$$

These equations represent the frictional boundaries for the thrust and normal faulting regime. The frictional boundary for the strike-slip faulting regime, where $\sigma_1 = \sigma_H$ and $\sigma_3 = \sigma_h$, can be calculated by substituting σ_v by σ_{hmin} in Eq. (5.6). Zoback et al. (2003) used these equations to determine the crust's frictional stability in a horizontal stress space and introduced the stress polygon, which allows to illustrate a possible range of σ_H and σ_h at depth z for a given pore pressure and a given friction coefficient (Figure 5.1a). In order to use these frictional limits together with our definition of the stresses, we substitute σ_v , σ_{Hmax} , σ_{hmin} in Eqs. 5.6 and 5.7 by Eqs 5.1, 5.4 and 5.3, respectively and solve them for ε_H to represent the crust's frictional stability polygon in the horizontal strain space (Figure 5.1b). Doing this conversion helps to better constrain possible values for ε_h and ε_H for our analytical stress model. In addition to this strain polygon, we also consider the criticality of the faults. At a particular depth, for each strain combination, we get a value for σ_h and σ_H . For a critically stressed fault, values of stresses must be such that the shear stress on the fault is close to the rupture point. To evaluate the distance of the shear stress τ on a given fault to the rupture point, we use the Coulomb failure stress (CFS), defined as follows (assuming a cohesionless environment) (see e.g. Beeler et al., 2000):

$$CFS = \tau - \mu(\sigma_n - p) \quad (5.8)$$

where p is pore pressure. A negative value resulting from Eq. (5.8) indicates that failure has not been reached, as the shear strength ($\mu(\sigma_n - p)$) is greater than the shear stress (τ), while a positive value indicates rupture. Repeating this for several critically stressed faults of different orientations, and assuming that they all rupture under similar stress conditions, allows to identify strain combinations that generate plausible stress conditions for all faults. These different strain combination are used to calculate stresses over depth using Eq. (5.1)-(5.4) and stress profiles are plotted.

Stress inversion

Stress inversion methods have first been developed to analyse stress and strain in calcite twinning (Weiss, 1954; Turner, 1962; Groshong, 1972; Burkhard, 1993), and later on for calculating the paleo-stress tensor from field observation of striations on fault surfaces (Angelier, 1979; Michael, 1984). They have then been extended to calculate present day stress field from focal mechanisms

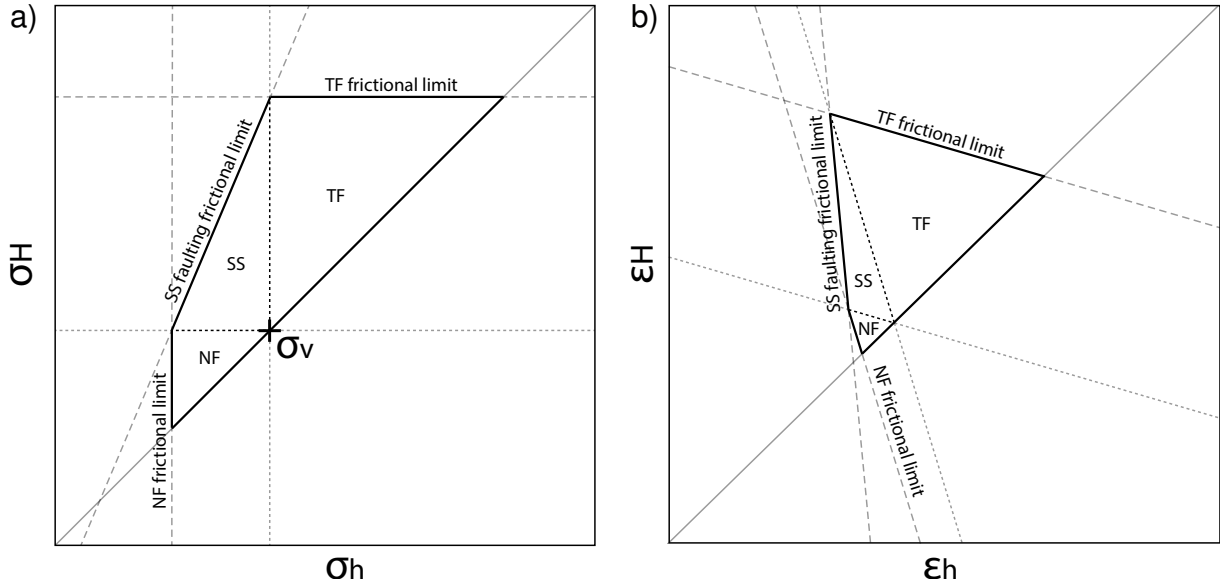


Figure 5.1: a) Stress polygon illustrating the crust's frictional stability boundaries in a stress space, as defined in [Zoback et al. \(2003\)](#) and b) Strain polygon illustrating the crust's frictional stability in a strain space. NF: Normal faulting, SS: Strike-slip faulting, TF: Thrust faulting.

([Michael, 1987](#); [Angelier, 2002](#); [Gephart and Forsyth, 1984](#)). Among the most widely used methods are the ones of [Angelier \(2002\)](#), [Michael \(1987\)](#) and [Gephart and Forsyth \(1984\)](#), which are all based on three common assumptions: (i) Faults slip in the same direction as the resolved shear stress, this is known as the Wallace-Bott hypothesis ([Wallace, 1951](#); [Bott, 1959](#)). (ii) Physical properties of the rock are assumed to be homogeneous and stress is assumed to be constant within the considered volume. (iii) Focal mechanisms must be independent in order to avoid a perturbed stress field resulting from a previous earthquake. The major drawback of these methods is that a choice must be made on which one of the two nodal planes is the fault plane. To overcome this, [Lund and Slunga \(1999\)](#) added a fault stability criterion to the [Gephart and Forsyth \(1984\)](#) method and identify the fault plane as the most unstable nodal plane under a given stress field. [Vavryčuk \(2014\)](#) modified the method of [Michael \(1987\)](#) in a similar way. These methods use a set of focal mechanisms to determine a single reduced stress tensor (optimal solution) valid for all fault planes by maximizing the objective function $\mathbf{F}$. Where $\mathbf{F}$ is a four-dimensional parameter space function (three angles and the relative principal stress ratio). The optimal solution is the maximum point of this function.

In this study, we follow the protocol of [Kakurina \(2019\)](#), inspired by the one of [Yamaji \(2003\)](#), which uses the inversion method of [Angelier \(1979\)](#), but also considers the topography of the objective function $\mathbf{F}$, by considering all the solutions (and not only the maximum). While the protocol of [Yamaji \(2003\)](#) is for a set of faults, the one of [Kakurina \(2019\)](#) is for single fracture or fault. A computational grid of the four-dimensional parameter space is calculated using the [Rakhmanov et al. \(1994\)](#) algorithm, that generates a mesh composed of 300 directions of σ_3 , 18 directions of σ_1 for each σ_3 , and 11 variations of the stress ratio R for each combination of σ_1 and σ_3 . In total $300 \times 18 \times 11 = 59400$ grid points are generated. The objective function $\mathbf{F}$ consists in maximizing the fit of the 59400 stress tensors from the grid to the observed data, in our case the

slip direction. $\mathbf{F} = f(d)$, where d is the angular misfit between the calculated and the observed slip direction for a given stress and $f(d)$ can be of any form as long as it is a decreasing function of d . In [Yamaji \(2003\)](#), the inversion is done for a set of faults and $\mathbf{F} = \sum_i^N f(d_i)$, where d_i is the misfit angle of the i th fault and N , the total number of faults. The angular misfit d can vary between 0° and 180° , but since the Wallace-Bott hypothesis assumes collinearity between the resolved shear stress and the slip direction, the misfit angle should not exceed 45° . This is also suggested by [Angelier \(1979\)](#) who recommends to suppress the influence of outliers with the following definition of $f(d)$:

$$f(d) = \begin{cases} 1 - \tan(d) & (0 \leq d \leq 45^\circ) \\ 0 & (45^\circ < d) \end{cases} \quad (5.9)$$

In our case, we set the maximum misfit angle at 5° to drastically reduce the initial number (59400) of possible stress tensors. Once the inversion is done, we further reduce the number of possible stress tensors by considering the principal stress magnitude and the slip tendency (definition follows). Principal stress magnitudes are calculated based on estimations of the vertical stress σ_v and the stress ratio by solving the following system of equations:

$$\begin{cases} R = (\sigma_2 - \sigma_3)/(\sigma_1 - \sigma_3) \\ \sigma_v = l^2\sigma_1 + m^2\sigma_2 + n^2\sigma_3 \end{cases} \quad (5.10)$$

where l , m and n are the cosines of the vertical component of the fault. Since we have 3 unknowns (σ_1 , σ_2 and σ_3) and only 2 equations, we add a third equation that states that the friction on the fault must be equal to 0.75. We then solve the system of the above equations as functions of σ_3 , and then numerically search for a value of σ_3 that satisfies all equations. The value of σ_3 must be higher than the pore pressure to avoid being in the tensile domain and lower than σ_v , as the minimum principal stress cannot be greater than the vertical stress component. Finally, for each possible stress tensor, the effective normal stress, $\sigma'_n = \sigma_n - p$ and the shear component τ are calculated to evaluate the slip tendency T_s , where $T_s = \tau/\sigma'_n$. Only stress tensors with T_s between 0.6 and 0.8 are kept as valid stress tensors. Note here, that p is the hydrostatic pore pressure and that the triggering pore pressure is not considered since it only has a minor effect.

5.3 Data

Fault plane orientations

The data used for the stress estimation of the Neuchâtel Jura consists in a collection of fault orientations, dips and rakes of the slip. For the analytical stress model, we use fault orientation from mapped structures, while we use the orientation of computed focal mechanisms for the inversion, either from this study or from literature ([Chapter 3](#)). Please note that the uncertainty in the composite fault plane solution identified in this study is high, as the quality ranking of earthquake location is low (see [Appendix F](#)), uncertainties in hypocenters can result in considerable different fault plane solutions. For the analytical stress model, we consider structures that are supposed to be critically stressed, i.e. that reach their rupture point following small pressure

Table 5.1: Fault plane orientations defined for the analytical calculation of stresses, based on local geological map, where fault az. represents the azimuth of the dip.

Name	Fault az.	Fault dip
Lessy	210	85
Neuveville	330	40
Ferrière	100	85
Saint-Blaise	89	90
La Lance	190	85

Table 5.2: Orientations of the nodal planes identified as the fault planes.

Name	Fault strike	Fault dip	Rake
F1	350	75	0
F3	126	82	-66
F4	310	90	160
F5	4	67	44
L1	355	86	-2
L4	250	40	128
L5	276	86	115
L6	228	60	6

variation induced by heavy rainfalls (Chapter 4), these structures are listed in Table 5.1. For the stress inversion we use all known focal mechanisms available within the study area that are located at depth < 5 km, to constrain the stress inversion to the upper crust (i.e. the sedimentary cover), and select the nodal plane according to interpretations in Chapter 3 (Table 5.2).

Parameters

In addition to the fault plane orientations, the depth at which the stresses are calculated and physical properties of the crust must be defined. We assume similar stress conditions for the entire study area, between 200 and 2000 m.b.s.l. The lower limit represents the transition between the sedimentary cover and the crystalline basement Sommaruga (1997), while the upper limit is set to avoid surface-related sources of stress (e.g. relaxation and non-elastic behaviour) as suggested in Zoback et al. (1989). Table 5.3, lists all the parameters and the chosen values used for the stress estimation. The chosen value for the Young's Modulus is representative for the stiffer formations, where the differential stress is the highest and where rupture is expected to occur first. For the stress inversion, the depth is fixed at 1000 m.b.s.l. for all focal mechanisms, which is an average and consistent for our focal mechanism given the large uncertainty on depth determination. For the analytical model, we assume that deformation is constant over depth, and that the chosen strain combination and the resulting stresses must equally satisfy the rupture conditions for near surface structures and deeper faults. In order to satisfy both, rupture taking place near the surface and further at depth, we attribute two distinct depths, 500 m.b.s.l and 2000 m.b.s.l to structures showing rather superficial or deeper deformation, respectively, in order to calculate the stresses. Moreover, based on local deformation structures like thrust faults and fold axis of the region and compilation of strain field data inverted from geodetic measurements

(Houlié et al., 2018), we set the orientation of the maximum horizontal strain ε_H to an azimuth of 135°N .

Table 5.3: Parameters definition for the analyzed structures. Average altitude at activated area (altitude at epicenters), piezometric level to calculate pore pressure and depth of interest to calculate the vertical stress. Values are in m.a.s.l. Second half of the table shows the values of physical parameters and conditions estimated from values found in literature (Hergert et al., 2015; Houlié et al., 2018) .

Name	Altitude	Piezometric level	Depth
F1	300	463	-1000
F3	485	467	-1000
F4	470	430	-1000
F5	1250	600	-1000
L1	300	463	-1000
L4	850	535	-1000
L5	1330	463	-1000
L6	790	600	-1000
Lessy	300	463	-500
La Neuveville	850	535	-2000
La Ferrière	1250	600	-2000
Saint-Blaise	465	460	-500
La Lance	750	460	-500
Input parameters			
Rock density ρ_r [kg/m ³]	2650		
Water density ρ_w [kg/m ³]	997		
Gravitational acc. g [m/s ²]	9.81		
Young's Modulus E [GPa]	40		
Poisson's ratio ν [-]	0.3		
Friction coefficient μ [-]	0.85		
Azimuth ε_H [°]	135		
Azimuth ε_h [°]	45		

5.4 Results

Analytical stress model

We build the strain polygon for different altitudes (i.e. 500 and 2000 m.b.s.l.) and plot the rupture line for each fault as the $\text{CFS} = 0$ contour line on the strain polygon plot corresponding to the focal depth defined earlier. Since we assume that faults must rupture under the same strain conditions, we select the most restrictive strain polygon to build our stress models (i.e. the strain polygon at 500 m.b.s.l.). We test 3 different models: **A**, **B**, **C** and **D**, corresponding to the corners of our strain polygon (Figure 5.2) and plot the stress profiles accordingly. Figure 5.3 shows the stress profiles for each strain combination at corners A,B,C and D. Models **A** and **B** result in minimum horizontal stresses that are lower than σ_{hmin} towards depth. Model **C** shows σ_h and σ_H being close to the minimum and maximum possible horizontal stresses, respectively. In model **D** $\sigma_h = \sigma_H$ and is close to the maximum possible value. For our stress model, we want strain combination resulting is stress conditions valid between altitudes of -200 to -2000 m,

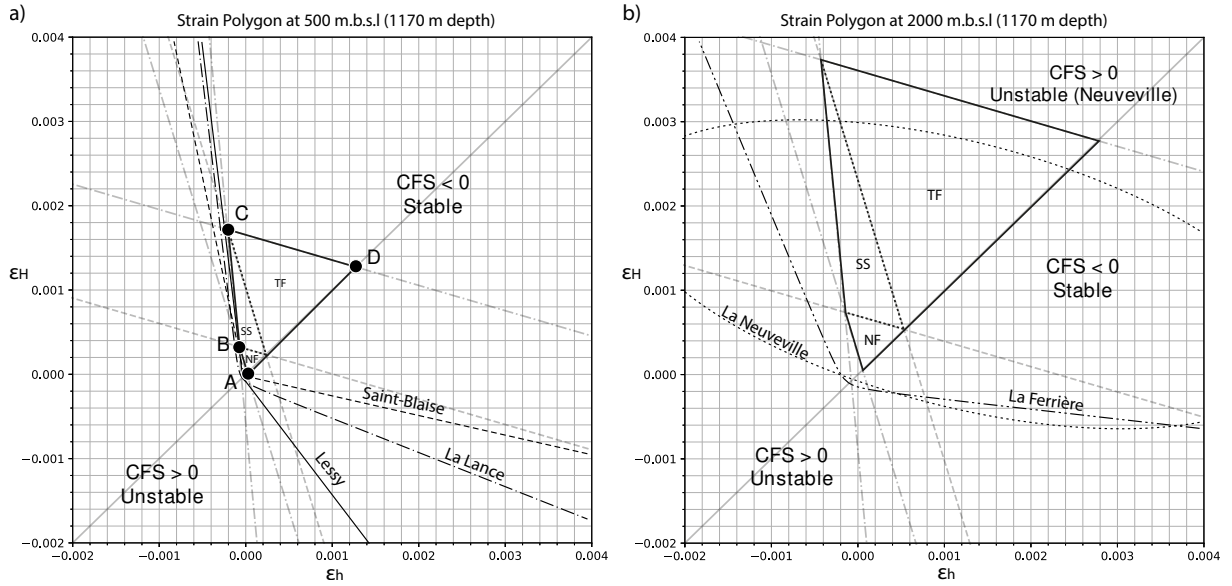


Figure 5.2: a) Strain polygon at altitude 500 m.b.s.l. and contour lines $CFS = 0$ corresponding to the rupture lines of different faults thought to have ruptured at this altitude, in the upper right corner of the plot, $CFS < 0$, and the fault is stable, while in the lower left corner $CFS > 0$ and the fault is unstable. Points A, B, C and D show the strain combination used for the models. b) same as a) for a altitude of 2000 m.b.s.l.

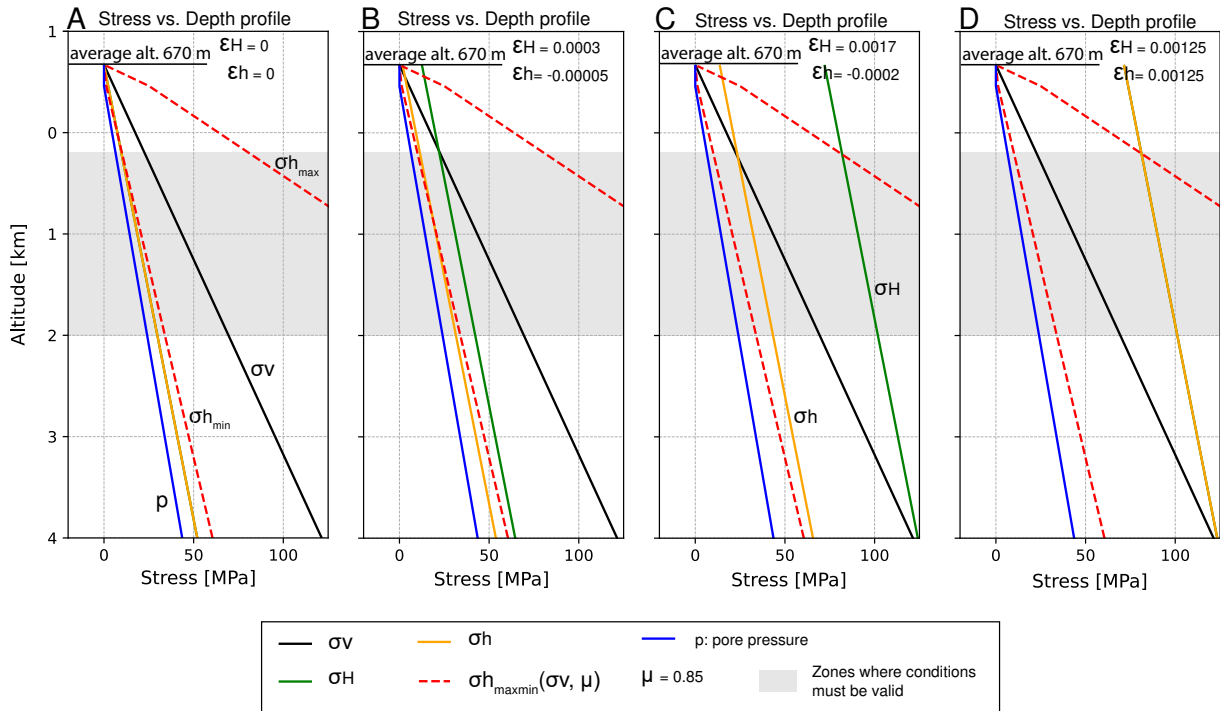


Figure 5.3: Stress profiles for different combinations of horizontal strains identified in Figure 5.2. Blue: hydrostatic pore pressure (p), yellow: minimum horizontal stress (σ_h), green: maximum horizontal stress (σ_H), black: vertical stress (σ_v), dashed red: minimum and maximum possible horizontal stress (σ_{hmin} and σ_{hmax}). Shaded area corresponds to the depth range within which the stress conditions must be valid, i.e. within the crust's frictional boundaries.

which means that models **A** and **B** are excluded. In addition we assume that σ_H must be greater than σ_h , thus model **D** is excluded. Model **C** fulfills all the conditions, and is kept as our best estimate analytical stress model.

Stress inversion from focal mechanisms

Following the protocol of [Kakurina \(2019\)](#), the results of the stress inversion are not unique, i.e. a single stress tensor is not determined for each fault. However, considering a maximum misfit angle (5°), upper and lower limits for σ_1 and σ_3 (Eq. (5.5)), and keeping only stress tensors generating a slip tendency between 0.6 and 0.8, the final number of solutions is considerably reduced in comparison to the number of initial solutions (59400). For all 8 fault plane solutions used for stress inversion, the number of final solutions is between 540 and 670. While the number of solutions is reduced, the spread of directions of possible stress tensors is still relatively high and explained by the use of single fault plane solution for the inversion. Looking at the results, σ_1 is mainly oriented NW-SE, between 290° and 350° with small plunge ($0-17^\circ$), except for F3 and L6, where σ_1 has a rather NE-SW trend. σ_3 is mostly located in the bottom left quadrant (between S and W) with a plunge that varies from close to 0° to more than 65° . Only L6 shows a SE-NW trend for σ_3 . For F1, F4, L1, L6, the closest stress to vertical is σ_2 , while it is σ_3 for F3, F5, L4 and L5. Figures. 5.4-5.5 show the final solutions of possible stress tensors and magnitudes (a), the mean value for each stress and the density contour of the solutions (b) and the distribution of stress magnitudes for σ_1 , σ_2 and σ_3 (c). Magnitude of stresses for σ_1 are relatively poorly constrained and the difference between values of the first and third quartile is between 40 and 80 MPa. For σ_2 and σ_3 , the differences is smaller in some cases, especially for solutions where one of the stresses is vertical, or near vertical, since the vertical stress magnitude is directly calculated from the overburden of the sedimentary layers. Thus, for solutions F1, F4, L1 and L6 for σ_2 and solutions F3, F5, L4 and L5 for σ_3 , the difference between values of the first and third quartile is around 10-15 MPa.

Stress inversion using STRESSINVERSE

The program developed by [Vavryčuk \(2014\)](#) allows a similar stress inversion but considers all fault planes together. Estimation of the stress magnitudes are not implemented in the program, but similar to the method of [Kakurina \(2019\)](#), the inversion is done for different stress ratios. The stress ratio in [Vavryčuk \(2014\)](#) is defined as:

$$R_v = \frac{\sigma_1 - \sigma_2}{\sigma_1 - \sigma_3} \quad (5.11)$$

where the index v stands for the authors name to avoid confusion with the stress ratio defined earlier (Eq. (5.10)). In addition to the stress inversion, the program calculates the slip tendency between nodal planes under a given stress state and selects the nodal plane with the highest slip tendency for the final inversion. In our case, 3 fault planes were inverted with their auxiliary plane (F3, F5 and L5). Results of the stress inversion are shown in Figure 5.6. Stress orientations are similar to the orientation found earlier, roughly NW-SE for σ_1 , SW-NE for σ_3 and almost vertical for σ_2 . Figure 5.6b shows the number of solutions for each stress ratio. Based on Eq. (5.11), R_v tends to 1 when σ_2 tends to σ_3 and R_v tends to 0 when σ_2 tends to σ_1 . Results suggest that for the majority of the solution, magnitudes of σ_2 are similar to σ_3 ($R=0.7-0.75$), when $\sigma_2 \approx \sigma_v$. For some solutions, magnitudes of σ_2 are closer to σ_1 ($R=0.35$), when $\sigma_3 \approx \sigma_v$.

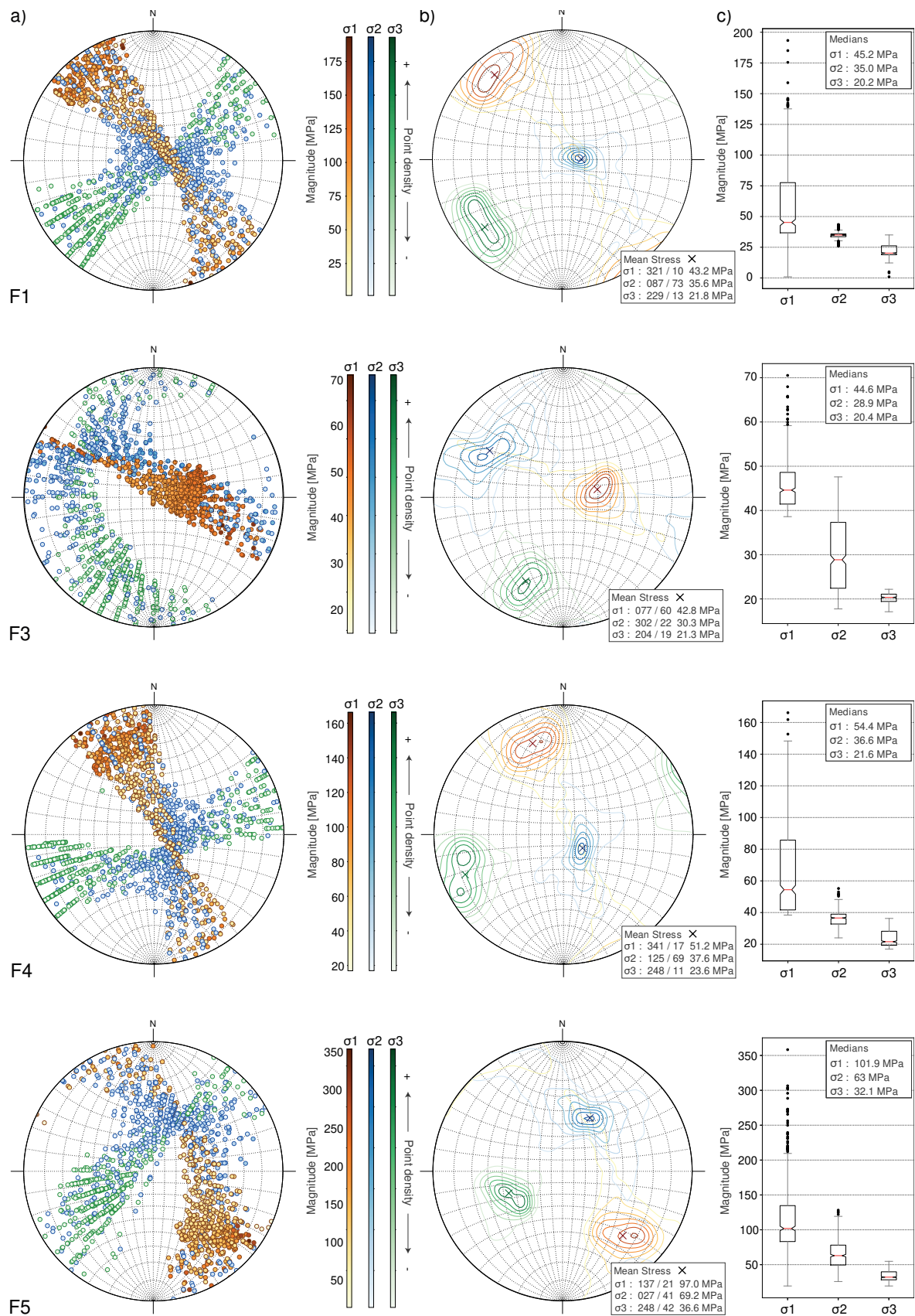


Figure 5.4: Stress inversion results for focal mechanisms F1, F3, F4 and F5. a) Stress tensors and magnitude, b) mean solution for each stress and point density and c) distribution of stress magnitudes.

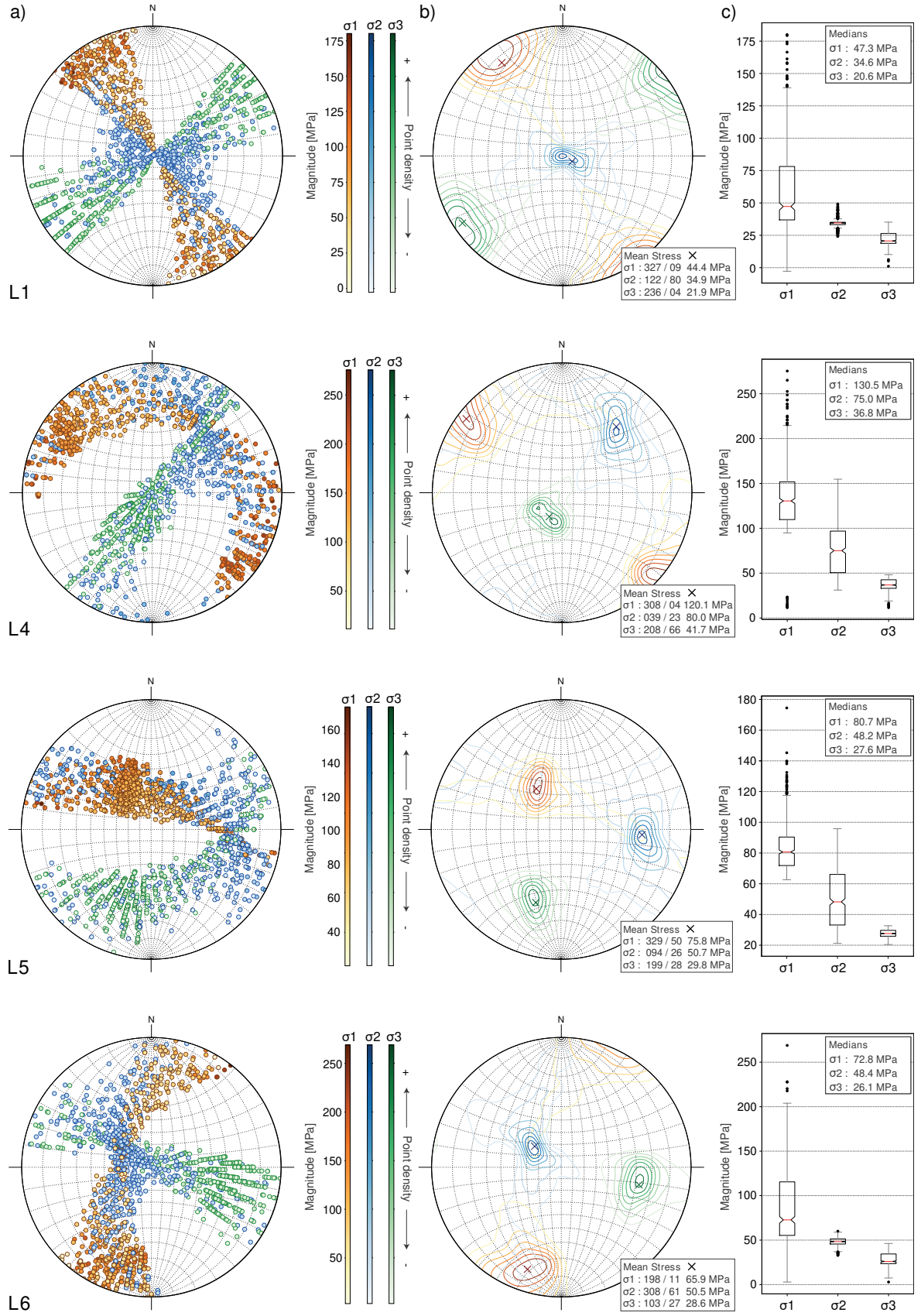


Figure 5.5: Stress inversion results for focal mechanisms L1, L4, L5 and L6. a) Stress tensors and magnitude, b) mean solution for each stress and point density and c) distribution of stress magnitudes.

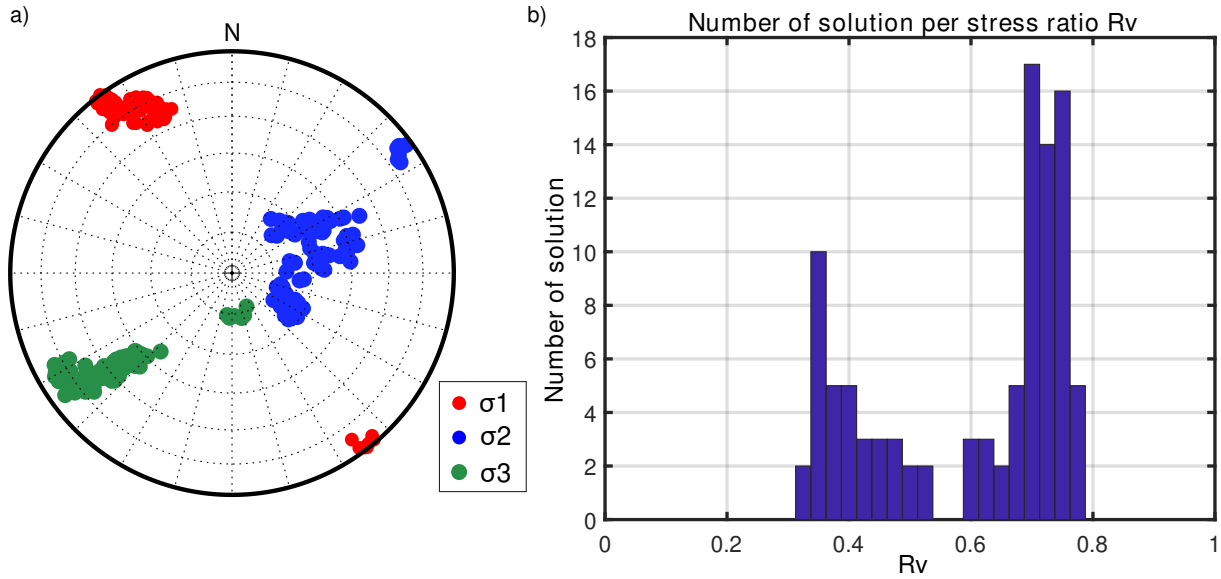


Figure 5.6: Distribution of direction of the 3 principal stress (a) for different values of stress ratios (b) resulting from the stress inversion using multiple focal mechanisms and the STRESS-INVERSE Program.

5.5 Discussion

Stress magnitudes and orientations

We compare results of our analytical stress model and the ones obtained from the inversion of fault plane solutions. While the stress orientations for the analytical stress model are defined, i.e. the maximum horizontal stress has an azimuth of 135°N (co-linear with the orientation of the maximum strain), the orientation of the stresses in the inversion is determined based on the stress ratio and the slip direction. The majority of the solutions show a NW-SE trend for σ_1 , a SE-NW trend for σ_3 and a vertical trend for σ_2 . Following the Anderson's definition of the stress regime (Anderson, 1951), these orientations correspond to a strike-slip environment, where the vertical stress is the intermediate stress. For the analytical stress model, the stress regime varies with depth and depends on the strain scenarios considered. For example, at 1000 m.b.s.l., the stress regime differs between scenarios **B** and **C**: normal faulting for **B** and strike-slip for **C**. From the stress inversion, we see that the majority of fault plane solutions show a strike-slip regime, i.e. F1, F4, L1 and L6, thus for comparison we plot the stresses of the inversion on model **C** (Figure 5.7). The depth of focal mechanisms has been set to 1 km.b.s.l., but the topographic elevation is taken into account in the inversion, thus values of mean stresses are plotted according to the total thickness between focal depth and the surface. Magnitudes of the minimum horizontal and vertical stress fall within the results of the analytical stress model and the mean stress values (circles) are close to the analytical model. Values for the mean maximum horizontal stress are much lower than the analytical model. Orientations are similar to the analytical stress model, except for L6 where the maximum horizontal stress, i.e. σ_1 in Figure 5.5 is oriented NNE-SSW. For the other fault plane solutions, i.e. F3, F5, L5 and L4, they show either coupled deformations (F3, F5 and L5) or a thrust regime (L4) and can therefore not

be represent on model **C**. For F3 the maximum principal stress is the closest to the vertical stress and the trend of the intermediate stress is close to the orientation of the maximum horizontal stress defined in the analytical stress model, suggesting a fault with a normal component. For F5, the trend of the maximum principal stress corresponds to the NW-SE orientation of the maximum horizontal stress. The intermediate and minimum stresses have a similar plunge, and none of them can be identified as the vertical stress suggesting a coupled reverse and strike slip movement. For L5, the maximum principal stress plunges 50° towards the NW, this orientation together with a E-W, nearly vertical fault plane (dip of 86° toward N), leads to a dominant reverse faulting.

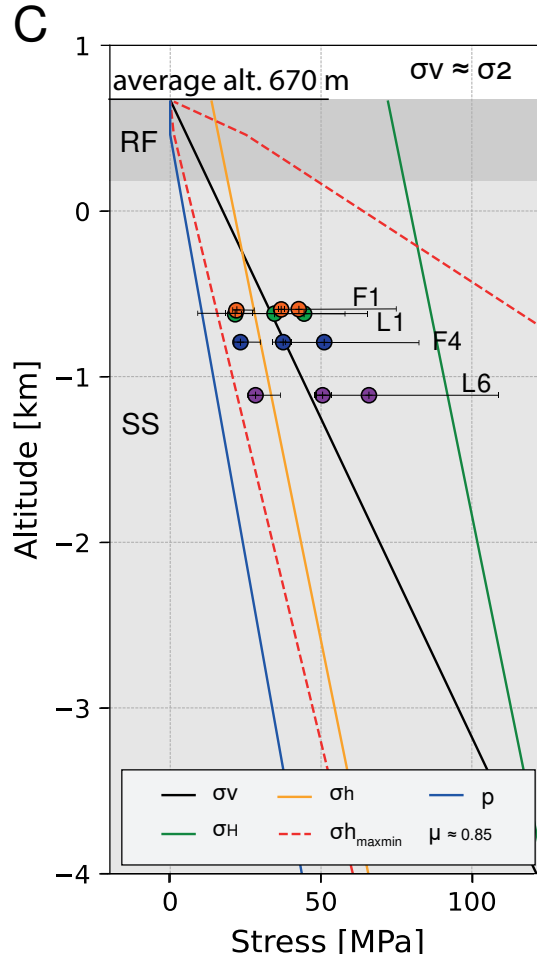


Figure 5.7: Results of the stress inversion for focal mechanisms F1, F4, L1 and L6 plotted on the analytical stress model **C** for comparison, circles represent the mean values and error bars are represented as the range of values (1^{st} and 4^{th} quartiles). Focal mechanisms F3, F5, L4 and L5 are not represented due to a different regime at focal focal depth.

Limitations of the models

Stress inversion

The stress estimation from the inversion of focal mechanisms is performed by setting values for the focal depth, the friction coefficient and the maximum misfit angle between the observed

and calculated slip orientation. Several tests have been performed by changing one or the other of these values: Stress orientations do not vary significantly when focal depth is changed, nor when the coefficient of friction is modified. There is a small change in the stress gradient with a varying friction; it increases slightly for a decreasing μ . The number of final solutions is obviously dependant on the misfit angle, and increases for an increasing angle, however, the density contour plot of the solutions and the medians of the stresses are similar. Since the inversion is based on the Wallace-Bott hypothesis, the major parameter affecting the estimation of the reduced stress tensor is the orientation of the fault plane and the slip direction, thus, a high uncertainty in the focal mechanisms and the identification of the fault plane leads to high uncertainties in the stress estimates. The stress inversion performed with STRESSINVERSE considers this uncertainty in the nodal plane selection and switches the planes during iteration to keep the nodal planes with the highest slip tendency under a given stress state. From the 8 fault plane solutions used for the inversion, 3 have been switched in our case (F3, F5 and L5). The nodal plane striking NW-SE was initially identified as the fault plane for F3, with a dominant normal faulting component: a coherent choice based on possible N-S striking structures identified on the digital terrain model (see Figure 5.10). For F5, the initial nodal plane identified as the fault plane is the one striking N-S, possibly corresponding to the Ferrière Fault, with a combined left lateral strike-slip and reverse component. The nodal plane striking E-W in L5 was selected as the fault plane since it corresponds approximately to the orientation of the La Lance fault and shows a right lateral oblique slip. After executing the STRESSINVERSE program, the nodal planes with the highest slip tendency and thus considered as the fault planes for the stress inversion are a NW-SE shallow dipping strike-slip fault for F3, a WNW-ESE right lateral strike-slip fault with a reverse component for F5 and a N-S shallow dipping left lateral strike-slip fault with a small reverse component. For F3 and L5, the newly identified fault planes are difficult to justify since no such structures are identified on the geological map nor have they been mentioned in literature. For F5 the strike and dip of the fault could correspond to a back thrust fault, a structure already identified further south (NW of Neuchâtel) and to the east (N of La Neuveville) and coherent with the general context of a principal horizontal stress oriented NW-SE. Although there is no certainty about the real fault plane, the initial choices are based on mapped structures, whereas the newly identified fault planes are based on the slip tendency. The slip tendency analysis tells us which nodal plane has the highest chance to slip, assuming both solutions are actual fault planes. Focal mechanisms tell us the strike and dip of the fault plane and gives us the auxiliary plane, but it does not mean that this plane is a fault. Thus, the slip will occur on the nodal plane which is an actual fault plane. Figure 5.8 shows the comparison of results from the stress inversion between the initial choice and the fault plane selected after executing STRESSINVERSE. For F3, the orientations and magnitudes are fairly similar, although results for the initial choice tends more to a normal faulting regime, in comparison to the STRESSINVERSE choice, with the mean orientation σ_1 being close to vertical. For F5, the trend of the plunge of the mean σ_1 has changed from the SE to the WNW, and for L5, the plunge of the mean σ_1 is smaller and slightly shifted towards the north. In general, magnitudes are much smaller for the stress inversion performed on the initial fault plane.

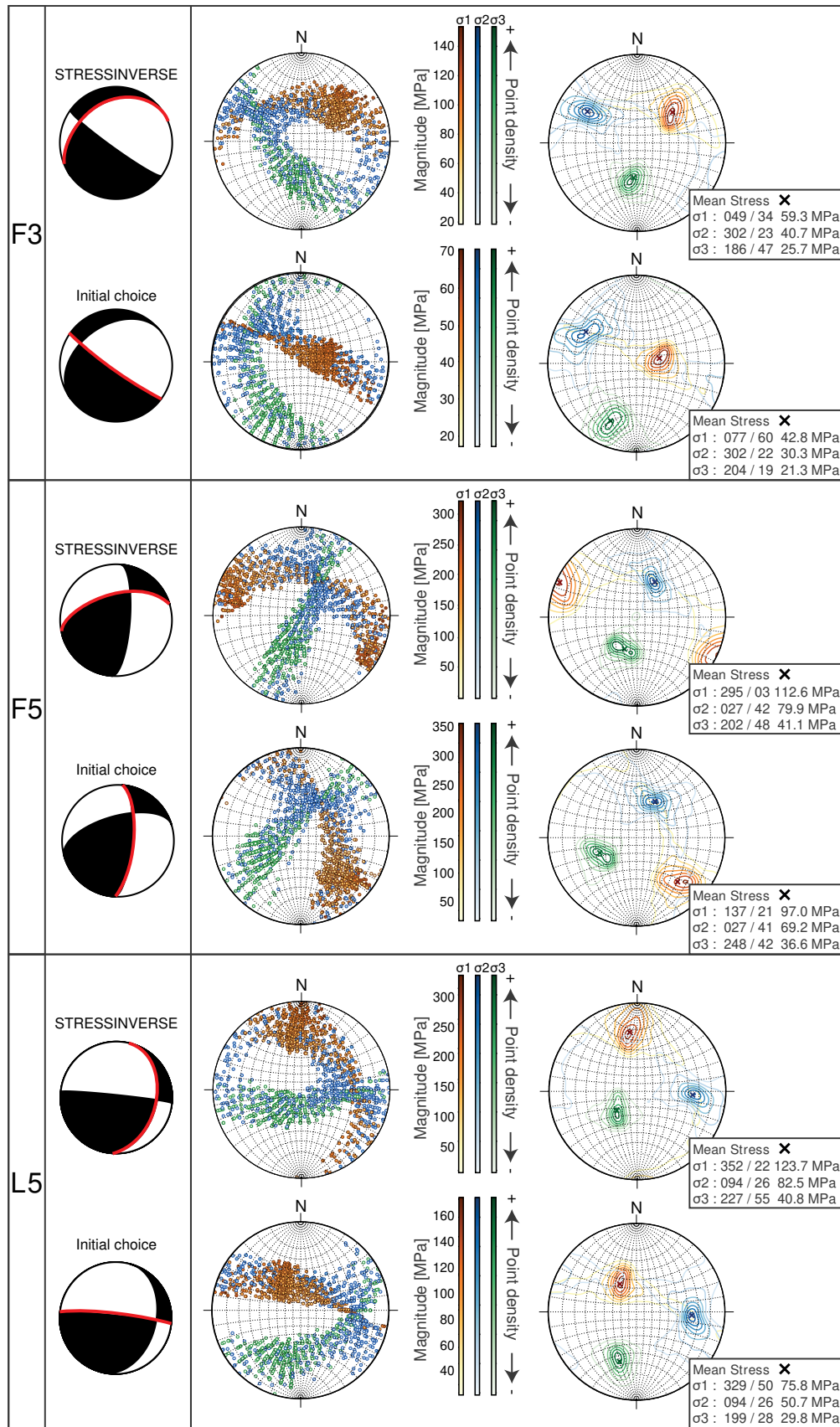


Figure 5.8: Comparison between results of the stress inversion using the nodal plane defined as the fault plane by STRESSINVERSE (top) and the one initially identified (bottom).

Analytical model

The analytical stress model presented here considers only the tectonic and gravity forces as major component of stress. Considering these two aspects, it could be improved by including geomechanical properties of the different layers, especially variations in the Young's Modulus. In their study, [Hergert et al. \(2015\)](#) show the major effect of the contrast of stiffness in a sedimentary sequence on the stress state under constant tectonic load. They conclude that the stiffer formations take up the majority of the tectonic stresses. This observation suggest that layers with the highest Young's Modulus accumulate more stress and are thus more likely to host seismicity under continuous tectonic loading. The value of 40 GPa used in our analysis and defined for the Upper Malm and Upper Muschelkalk units in [Hergert et al. \(2015\)](#), is taken for the entire depth in this analysis, assuming that seismicity occurs in the stiffer layers. Stress estimation using the highest Young's modulus, implies that values from the analytical stress model represent the upper bound of possible values. A lower Young's Modulus would result in smaller stress magnitudes. In addition to impacting the stress magnitudes over depth, [Figure 5.9](#) shows how changes in the elastic properties between layers of sediments can influence the stress regime. Better constraining the stress estimation at focal depth by considering the varying Young's Modulus would necessitate a higher accuracy in hypocenter location in order to determine with precision the sedimentary layer hosting the slip (i.e consider only earthquakes with a high-quality location ranking). Similarly to the material's stiffness, a variation of the Poisson's ratio over depth also influences the horizontal stresses. The value used in our analysis (0.3) corresponds to the highest estimate for the sedimentary units of the Jura Mountains ([Hergert et al., 2015](#)), thus calculated stresses with a lower Poisson's ratio would result in smaller magnitudes of stress. Although the analytical stress model is far from considering all aspects influencing the stress state, it does give a first estimate of possible stress magnitudes that are more or less coherent with results from the stress inversion. In addition to the aforementioned geomechanical properties to be adjusted for higher accuracy, a better knowledge about the actual strain would also help to calibrate the analytical stress model.

Structural analysis, stress map and slip tendency

Following Anderson's definition of stress regime, faults have been classified according to the direction of the mean stresses presented in [Figures 5.4-5.5](#). For F1, F4, L1 and L6, the intermediate principal stress is almost vertical, co-planar with the fault dip, suggesting a pure strike slip environment. For L4, it is the minimum principal stress that is the closest to the vertical stress, suggesting reverse faulting. For L5, we choose the solutions of the nodal plane initially identified as the fault plane (i.e. subvertical E-W striking plane) showing a reverse right lateral displacement. For F3, the solution of the initial choice is also preferred with a relatively steep dipping NW-SE striking fault showing a normal left lateral displacement. For F5 the proximity of the Ferrière fault tends to prioritize the initial choice of a near vertical dipping N-S striking fault with a reverse left lateral strike slip displacement. On [Figure 5.10](#), we propose a synthesis of the results from the stress estimation realized in this chapter. The orientation of the maximum horizontal stress tends to be NW-SE for the majority of the analyzed fault plane solutions, except for L6 where the orientation is NNE-SSW. The majority of the fault show a strike-slip or a cou-

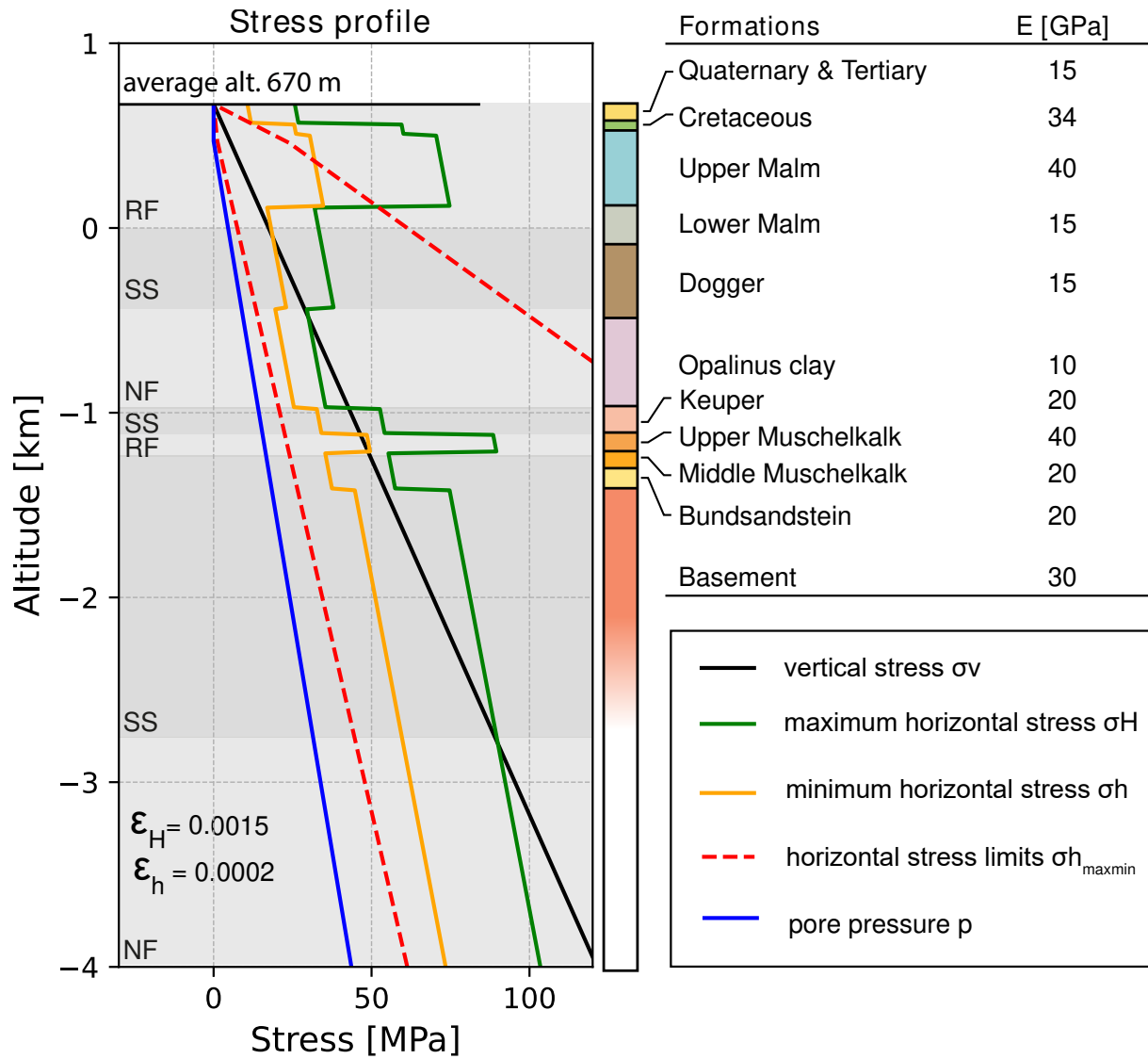


Figure 5.9: Stress profile over depth calculated using the analytical stress model with maximum and minimum horizontal strains of 0.0015 and 0.0002, respectively and varying Young's Modulus (E) determined for each layer (data from [Hergert et al. \(2015\)](#)). The stress regimes varies over depth (RF: Reverse faulting, NF: Normal faulting, SS: Strike-slip faulting) due to a varying Young's Modulus. The model aims at illustrating the high heterogeneity of the stress state and regime over depth. The stratigraphic column is considered as representative enough for the purpose of the figure, but does not consider any decollements levels which would induce strain partitioning and possibly even more heterogeneities.

pled reverse and strike-slip regime, except for L4 and F3, showing a purely reverse and coupled normal to strike-slip regime, respectively, suggesting that structural or lithological heterogeneities influence the stress state locally. Stress inversions analysis in the Jura Mountains were carried out in several studies in an attempt to better characterize the contemporary deformation style of this area (see for example [Heidbach and Reinecker \(2013\)](#)). [Becker and Werner \(1995\)](#) and [Becker \(2000\)](#) performed and analyzed in situ stress measurements in boreholes across the French and Swiss Jura Mountains. Results of these measurements for the Central Jura, corresponding to our study area, show a dominant N-S striking maximum horizontal stress near surface. They compared their results with fault plane solutions of earthquakes in the Jura Mountains showing a NW-SE orientation of the maximum horizontal stress and suggested an active stress decoupling between the basement and the cover. [Rabin et al. \(2018\)](#) used focal mechanisms to estimate the stress field. Their result for the Central Jura is relatively poorly constrained due to the low number of seismic events, however, an approximate NW-SE orientation for σ_H is still observed. Although they recognize that locating with precision the earthquake is difficult due to hypocentral error, they propose that the general stress field over the Jura Mountains has a σ_H oriented NW-SE from the basement towards the surface. [Lanza et al. \(2022\)](#) suggested a present-day stress field with a maximum horizontal stress oriented NW-SE in the basement but N-S in the sedimentary cover for the Northern Jura, however, authors recognized that in the Central Jura, near Neuchâtel, this anticlockwise stress rotation from the basement to the cover is not observed. In addition to the stress inversion of focal mechanisms and in situ measurements, deformation data at the surface from GPS networks revealed an extensional strain axis oriented NW-SE in Central Jura ([Houlié et al., 2018](#)). Authors interpreted this observation as the result of the motion associated to NW-SE strike-slip faults. While results of large scale studies tend to a NW-SE orientation of σ_H ([Rabin et al., 2018](#)), similar to our analysis, in situ near surface stress measurements tend to a N-S orientation of σ_H ([Becker, 2000](#)). Although we estimated the fault plane solutions used for the inversion to be located in the sedimentary cover, it is difficult to unambiguously attribute our stress field orientation to the upper crust because of the large uncertainty regarding focal depth, and thus impossible to confirm or infirm interpretations made by others regarding geomechanical coupling or decoupling between the basement and the cover. Nonetheless, our data contribute to filling the lack of focal mechanisms in Central Jura, and could be reassessed in the future using high precision relocation techniques in order to better constrain the depth. Moreover, although the general trend tends to a NW-SE orientation of σ_H local variations suggest that the stress field can rotate several degrees over a relatively short distance (e.g. L6 and F5 or F1 and L5) emphasizing the importance of considering the scale when comparing measurements and the presence of local stress disturbances (i.e. topography, rock anisotropy, dip of the layers ([Zang and Stephansson, 2010](#))). This is also true over depth as attested by the N-S oriented σ_H for measurements near the surface (tens of meters) ([Becker, 2000](#)) in comparison to our results (thought to be located between 0-4 km), where surface effects such as topography do not influence the stress field in the same manner. We know from microseismic monitoring (Chapter [3](#)) that the major faults in the study area show seismicity, and are thus preferentially oriented for a given stress tensor, but the depth at which they slip is still not well constrained. For a given stress field, we evaluate the tendency of the faults to slip over

depth. We test 2 different orientation for σ_H , (i) N-S and (ii) NW-SE, representing the 2 major orientations of σ_H found in literature and use the stress profile of model **C** to calculate stresses over depth, between 0.2 and 2 km.b.s.l.. We use the geological model of [Guglielmetti et al. \(2013\)](#) and the vertical to sub-vertical structures therein (i.e. no thrust faults). Fault's surfaces are modelled as triangulated surfaces, each node (triangle's vertex) is attributed an orientation based on the orientation of neighbouring surfaces, for which the shear and normal stresses and the ratio of both are calculated (i.e. τ/σ_n , the slip tendency). For visualizing purposes, the surfaces of the triangles are colored based on an interpolation between vertices. Figure [5.11](#) shows results of the slip tendency analysis over depth for the two tested orientations of σ_H . The slip tendency is the highest for all investigated faults when σ_H is oriented NW-SE, although the northern and upper parts of the Ferrière and Neuchâtel faults also show a high slip tendency when σ_H is oriented N-S. We suggest that the dominant NW-SE oriented σ_H stress field is more likely to trigger a slip on vertical to subvertical faults in the study area, mainly around the La Lance and Lessy Faults, which are also the structures showing the highest seismic rate. However local variation in the stress field might also generate a slip near surface on structures oriented similarly to the Ferrière and Neuchâtel Faults, which is coherent with the event L6, that has been located at a relatively shallow depth (altitude of $-0.86 \text{ km} \pm 1.6 \text{ km}$ according to [Abednego \(2015\)](#)). The stress field rotation from NW-SE to N-S can either be the results of local heterogeneities or it could indicate a regional anticlockwise stress rotation from the depth towards the surface, as mentioned by others (e.g. [Becker \(2000\)](#); [Lanza et al. \(2022\)](#)), assuming events showing a NW-SE orientation of σ_H are deeper than L6. This interpretation would be consistent with the near surface stress field found in [Becker \(2000\)](#), however to corroborate this hypothesis, high precision relocation of events used in this analysis would be necessary.

5.6 Conclusion

The aim of this analysis was to obtain a first estimation of the stress field in the study area by integrating knowledge about fault criticality, local tectonic setting and seismic activity, using 2 distinct methods. First the analytical stress model, considering gravity and tectonic forces as major component of the stress state, frictional boundaries and a series of conditions allowed to identify a plausible stress model. Second, the stress inversion from focal mechanisms revealed a range of possible stress tensors for each fault plane, with a NW-SE dominant trend for σ_1 and a dominant vertical orientation for σ_2 , suggesting that the majority of the investigated faults are under a strike-slip regime, although some results suggest a coupled regime between strike-slip and reverse ($\sigma_3 \approx \sigma_v$) or strike-slip and normal ($\sigma_1 \approx \sigma_v$). Only 1 fault plane solution shows a pure reverse faulting and is coherent with the presence of a back-thrust fault mapped near the epicenter. Orientations obtained from the inversion of focal mechanisms are relatively similar to the one defined in the analytical stress model and are in good agreement with existing studies on the stress orientation in the Central Jura. Magnitudes of stresses between the inversion and the analytical models are compared when appropriate, i.e. when the stress regimes found from the stress inversion are close enough to the stress regimes identified in the analytical model. For focal mechanisms showing a strike-slip regime, mean magnitudes of σ_3 and σ_2 , respectively σ_h and σ_v , are close to the values of the analytical model used for comparison (i.e. model **C**),

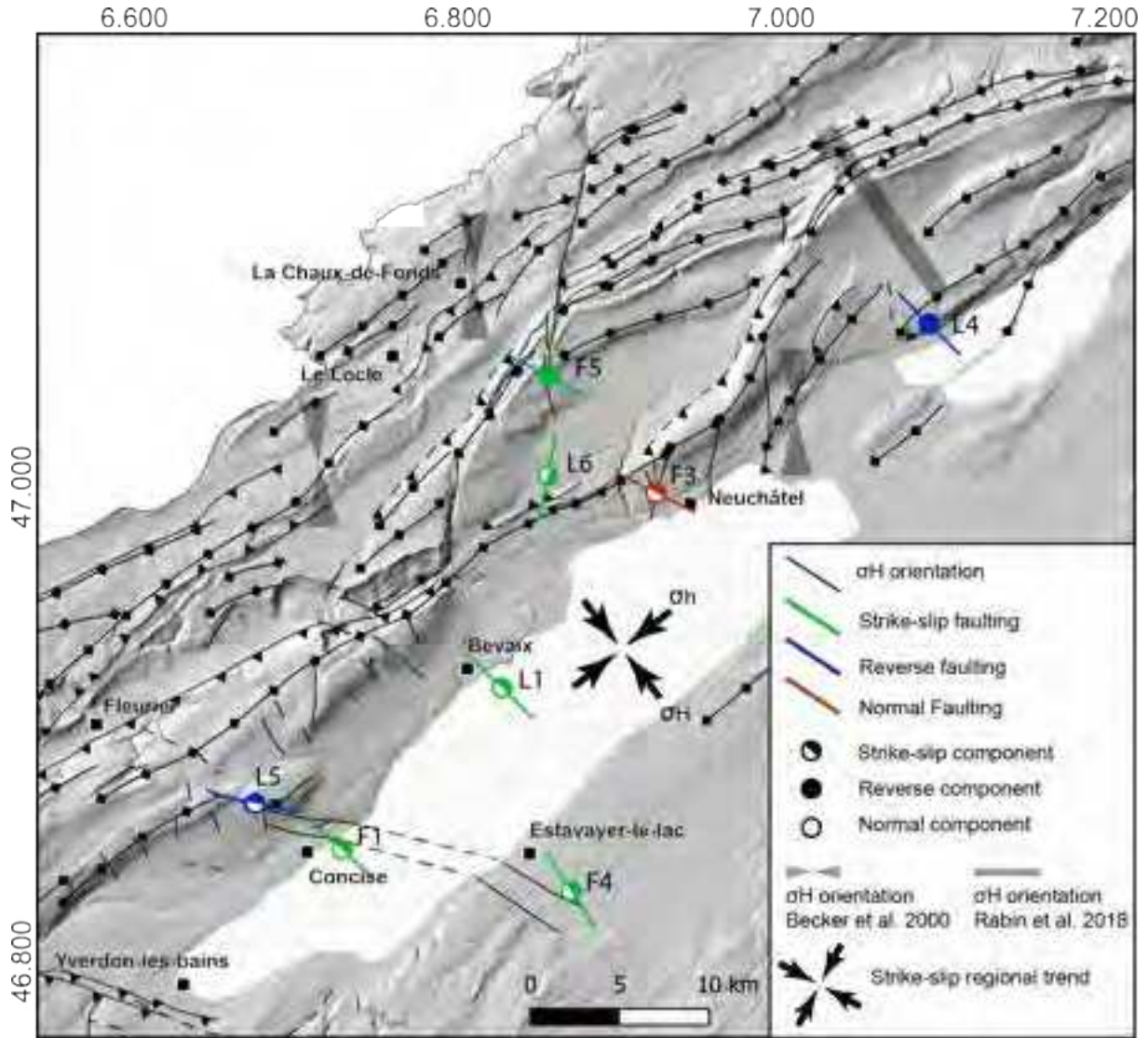


Figure 5.10: Regional stress map resulting from the inversion of focal mechanisms (F1, F3, F4, F5, L1, L4, L5 and L6). Also shown stress orientation of previous studies (Becker, 2000; Rabin et al., 2018)

while values for mean σ_1 , respectively σ_H , are much smaller. While uncertainty on the stress estimate from the stress inversion mainly lies in the focal mechanism itself and the identification of the fault plane, the analytical stress model is oversimplified by estimating a constant strain over the study area (horizontally and vertically) and constant elastic properties. Adjusting these parameters would result in a higher accuracy of the stress profiles, and eventually lead to a single stress profile satisfying every observations. The general stress field shows a NW-SE orientation of σ_H and a strike-slip regime, local variation suggest that the stress field can rotate over short distances. Although such variation might be due to local heterogeneities, a slip tendency analysis showed that for 2 different orientations of σ_H , N-S and NW-SE, the highest slip tendency for a given fault is found at different depth, suggesting that a varying stress field over a short distance (i.e. on the same fault segment) might be the result of a stress field rotation over depth. However, to unambiguously determine whether the local stress variation are due to

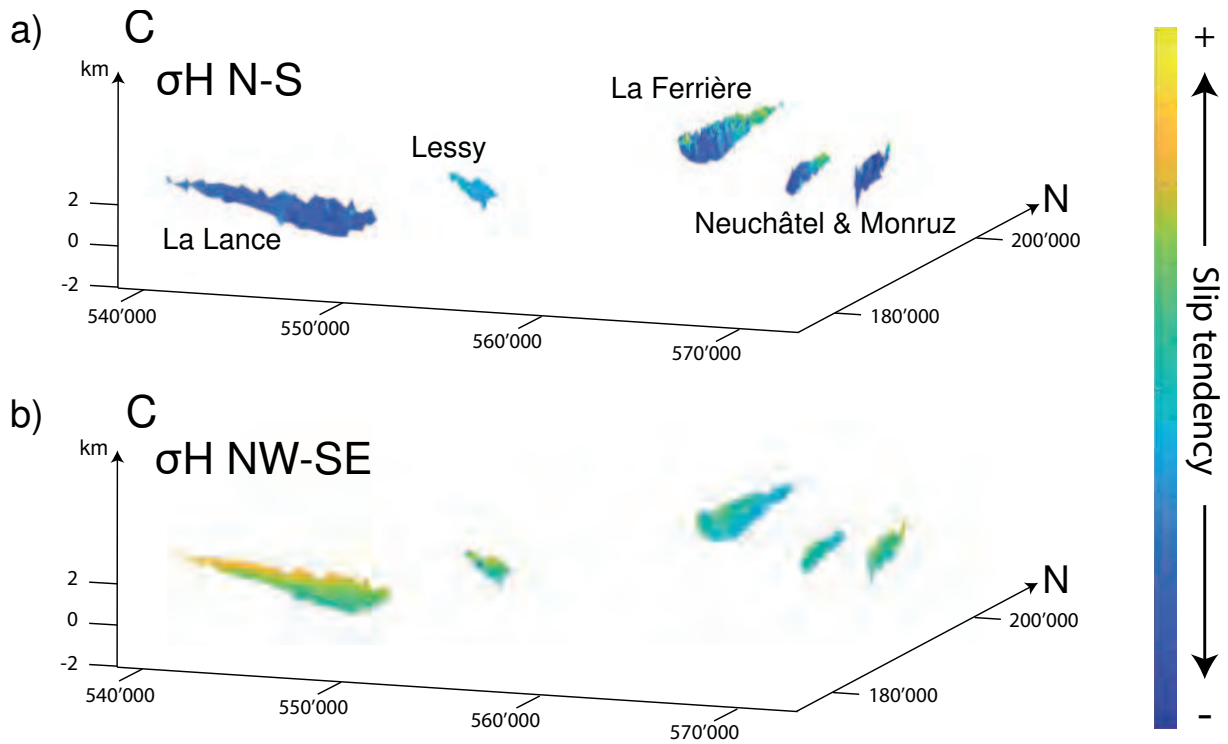


Figure 5.11: Slip tendency analysis over depth on investigated faults for a stress field with σ_H oriented N-S (a) and a stress field with σ_H oriented NW-SE (b) for stress model **C**

local heterogeneities or to a regional stress field rotation would necessitate a higher precision in the earthquake location and additional in-situ stress measurements from deep boreholes.

Chapter 6

Conclusion

6.1 Synthesis

The objectives of this thesis were to evaluate fault criticality in karstic regions using hydrogeological and microseismic data, to contribute to a better understanding of seismicity and its relation to groundwater circulation in the study area and use this knowledge to estimate the present-day state of stress of the shallow crust. These objectives were set based on current questioning regarding induced seismicity and its associated risks in the initiation of projects of deep geothermal energy, CO₂ sequestration, or radioactive wastes deposits, where the comprehension of the geological, hydrogeological and geomechanical context is essential.

In order to meet these objectives, the first part of the work consisted in identifying and quantifying the mechanisms that can trigger earthquakes in karstic regions following heavy rainfall. The mechanisms identified are: 1) an increasing vertical load, where the water filling the karstic conduits acts as an supplementary load, 2) the poro-elastic deformation induced by this same load, 3) the direct hydraulic connection between freshly infiltrated water and deep groundwater via the karstic conduits and fractures and 4) the diffusion of pressure from the base of the karstic conduits towards depth. These mechanisms, alone or in combination, act on the pore pressure, which, as it increases, reduces the stability of a fault by decreasing the effective stresses. Using 1D analytical solutions, we were able to determine that an increasing vertical load coupled with poro-elastic deformation only induce pore pressure changes of the order of kPa, whereas direct connection and pressure diffusion result in pore pressure changes of up to MPa. Although these mechanisms all act on pore pressure, the time at which they are effective varies: when the water level in the karst rises massively, the pore pressure response is direct in the case of an increase in vertical load and associated poro-elastic deformation, as well as for direct connection, but is delayed for pressure diffusion, and depends on the diffusivity of the medium as well as the distance the diffusion front travels.

By applying these solutions to 3 case studies, in which earthquakes occurred following heavy rainfall and considering the time lag between maximum rainfall and the onset of seismicity, we were able to identify which mechanism is more likely to generate earthquakes in karstic regions following a significant rise in groundwater level. In the cases of France, in the Avignon region, and central Switzerland, at Muotathal, seismicity occurred a few hours after the end of extreme precipitation, suggesting a mechanism with a direct effect on pore pressure. In

Germany, at Hochstaufen, seismicity starts about a day after precipitation. In the case of France and Switzerland, given the low magnitude of changes induced by increased vertical load and poro-elastic deformation, direct hydrogeological connection between surface and focal depth is the preferred mechanism. In the case of Hochstaufen, a coupled solution between the direct hydrogeological connection between the surface and the base of the karst system, potentially located several hundred meters below the surface, followed by pressure diffusion is considered a possible scenario explaining the delay between maximum precipitation and seismicity. Although the time between maximum precipitation and the onset of seismicity is an interesting indication, potentially informing us as to which mechanism is favoured, the timing of maximum precipitation intensities necessarily precedes the rise in water level in karst conduits. Thus, depending on the vegetation and sedimentary cover, infiltration may occur more or less rapidly, and the time lag between precipitation and seismicity is no longer representative of the mechanism involved. In order to compensate for this infiltration factor, which can vary in space and time, we propose that the time delay to be considered should be that between the moment of the massive increase in water level in the conduits and the onset of seismicity. Moreover, we advocate that the massive hydraulic head increase within karst conduits following heavy precipitation and the extend of the network at depth are the major factors influencing pore pressure variation at focal depth, and must therefore be considered when modelling pore pressure diffusion.

After this first part, which enabled us to identify the mechanisms more likely to generate earthquakes in karstic regions following heavy rainfall, we wanted to determine whether the faults in the Lake Neuchâtel region are also prone to rupture when the water level in the karstic conduits increases. Two monitoring networks, one seismic and the other hydrogeologic, were deployed over a 2-year period. For the first, around twenty geophones were installed near known faults to record their seismic activity. For the second, as the installation of measurement systems in karst conduits requires in-depth knowledge of speleology, it was decided to estimate the variations in groundwater levels by measuring the continuous flow at major karstic springs in the region. In addition to these flow measurements, a measuring point in a karstic sinkhole (Torrent sinkhole) was generously shared with us by the Spéléo Club des Montagnes Neuchâteloises. This measuring point gives us a direct indication of the increase in the hydraulic head during floods.

Processing and analysis of this data has produced a seismic catalogue of 129 localised events and 106 additional events detected using template matching. Most of the earthquakes are located close to known faults, and the majority of events lie to the south of the first Jura anticline, beneath Lake Neuchâtel. The hydrogeological data provided a continuous time series of flow rate variations over two years, indicating periods of flooding or drought, and thus reflecting the hydraulic head in the karstic massif. An initial analysis of these two sets of data showed that there is a link between the increase in groundwater level in the karst and the presence of earthquakes.

A more detailed analysis of seismicity near fault zones and of events occurring during flood periods was then carried out. Using hydraulic head data from the Torrent sinkhole and analytical solutions for the mechanisms identified in the first part of the work, it was possible to calculate the evolution of pore pressure during floods and to define an increase in pore pressure triggering a rupture for each fault zone. These pressures were used to assess the criticality of the faults, most

of which appear to be sensitive to pressure changes between 0.3 and 0.8 MPa. The results were then discussed in a more general context of the tendency for faults to slip under known stresses. Discussion about the results also include the irregularity of seismicity regarding fluctuating groundwater levels, as faults do not rupture every time the pore pressure increases. Reasons for this include the definition of rain-triggered events, the threshold detection of seismicity and other parameters regarding the fault's properties and rupture cycle. Finally, the distribution of earthquakes, mainly located beneath Lake Neuchâtel, was also discussed, revealing that the hydrogeological context of the study area probably has a major influence on the location of earthquakes, and that events, when occurring during flood periods, may indicate preferential groundwater flow paths.

The final part of the work consisted in an assessment of the state of stress in the study area. A first method, based on the main sources of stress in the crust, i.e. gravity and tectonic forces, was used for the estimation. This method requires knowledge about strains in the directions of the principal horizontal stresses, information that is difficult to acquire. Therefore, the crust's frictional stability boundaries are presented in a horizontal strain space (strain polygon), which allows to visualize the possible strain combinations to calculate stresses. In addition to this strain polygon, the rupture lines of the faults are shown in the same space, allowing to identify which strain combination leads to a stress state close to the rupture conditions for a given fault. A second method, using focal mechanisms and considering that the direction of slip is parallel to the shear stress on the fault plane, was used. The results show a dominant NW-SE oriented σ_H and a general strike-slip regime, sometimes coupled with reverse or normal faulting. Finally, we discuss how our results are influenced by the model's limitations, e.g. fault plane identification and rock heterogeneities, and interpret them with regard to existing data.

6.2 Major outcomes

Investigating the relationship between earthquake occurrence and heavy rainfall by combining hydrogeological data together with microseismic observations allowed to acquire a better understanding of what possible mechanisms might trigger seismicity following important precipitations and how the geological and hydrogeological setting affects pore pressure propagation. Field investigations improved the knowledge about hydrotectonics of the study area and the criticality of analyzed faults. In addition, an estimation of the current stress field in the study area was realized based on acquired data and knowledge of local geological conditions. As a summary, this study lead to the following outcomes and conclusions :

- 1) Karstic conduits and their extend towards depth must be considered when discussing pore pressure variation at focal depth. Not considering them leads to erroneous triggering pore pressure estimates of up to 3 orders of magnitude.
- 2) The seismic catalogue of the study area indicates zones of higher and lower activity. La Lance Fault is clearly the most active structure of the area regarding the number of events, followed by the Lessy Fault. Other zones, such as the Neuchâtel area and the southern part of the Ferrière Fault also show some activity, but in a more sporadic manner.

- 3) Seismic distribution can be explained by the geological and hydrogeological context. With layers dipping to the southeast, fluid flow is concentrated below the lake and an excess in pore pressure is observed (artesian well near Bevaix). The topographic variation (1000 m) between the summit of the first anticline (Mont Aubert) and the bottom of the lake implies a deficit in vertical stress below the lake. Together, these observation suggest that the effective stresses below the lake are smaller and favor rupture in this area.
- 4) Combination of hydrogeological data and microseismic observation suggest that earthquake are approximately two times more likely to occur during very wet periods. Collected data also suggest there is a 80% chance to detect seismicity when the groundwater level fluctuation in the Torrent sinkhole overcomes the 100 m threshold. However, these observations are based on only 2 years of data and would require a longer monitoring period to be confirmed.
- 5) Rain-triggered events occurring in the vicinity of known faults suggest that they are sensitive to an increasing pore pressure. The triggering pore pressure varies between 0.3 and 0.8 MPa for the different faults. But seismicity also occurs outside periods of higher groundwater level, indicating that the rupture is not solely influenced by the increasing pore pressure.
- 6) The location of rain-triggered events informs on the groundwater circulation, which is concentrated in aquifer layers and along faults.
- 7) An alternative methods to illustrate the crust's frictional stability boundaries is presented, as a function of the horizontal strains. This allows to better constrain the strain when calculating stresses using the simplified analytical models.
- 8) The orientation σ_H is dominantly NW-SE with local variation showing a N-S orientations. These observations are in good agreement with existing studies and possibly indicate an anticlockwise stress rotation from the depth towards the surface. A first estimation of magnitudes can be read from the stress profile, although the uncertainty is relatively high, especially for σ_1 .

6.3 Perspectives

The investigation of hydrotectonics and fault criticality in karstic regions performed in this study relies on a multidisciplinary approach, including hydrogeologic, seismic, and modelling methods, each of which can be improved its own way. Additional observation points of the hydraulic head within karst conduits could improve the estimation of the timing and the amplitude of pressure changes at depth following important precipitations, which would influence the identification of rain-triggered events, as it relies on these two parameters. Ideally, each fault zone should have its own hydrogeological observation point to insure reliable hydraulic head increases. These could be acquired by installing level logger within karstic conduits. In addition, a deeper knowledge of the aquifer's properties at greater depth, for example the Malm aquifer on the southern shore of Lake Neuchâtel, could be acquired by a deep borehole. Such infrastructures would allow a comparison

between the hydraulic head variation near surface (at the Torrent sinkhole for example) and the hydraulic pressure increase at depth. Even though the borehole might not reflect directly the hydraulic gradient within the conduits (if the borehole does not intersect karstic galleries) it could improve the estimation of the diffusivity within the aquifer and thus better constrain the triggering pore pressure. More advanced seismic investigation, using additional methods for a high-precision relocation would decrease the uncertainty regarding focal depth. This could be beneficial for the subsequent analysis of (i) defining the triggering pore pressure at focal depth using our simplified analytical solutions and (ii) better constrain the stress model. Analytical solutions for our model could be improved by including more realistic properties of the crust (i.e. changes in lithologies and geomechanical properties) and considering a multi-dimensional analysis. This would on one hand, reduce the uncertainty of the triggering pore pressure defined for a fault and thus help to better constrain its sensitivity to an increasing hydraulic head and on the other hand, improve the stress models and acquire a more accurate estimation of the stress field over depth. Nonetheless, although several aspects can be improved, the methodology used in this study allows a rapid assessment of whether or not a fault zone in a karstic environment is prone to fail following important hydraulic pressure changes. It is thus a first step in determining whether a region can host projects for which fault criticality is a major concern.

Hydrogeological study

Detailed water balance of the study area and its sub-compartment and repeated tracer testing under different hydrogeological conditions should allow to identify hydrotectonic compartments, determine their extent and eventually inform about their hydraulic properties. A special focus should be made on the eastern side of the study area, i.e. The Chaumont anticline, where the hydrogeological connections are not well constrained and where the hydraulic role of the Monruz and Neuchâtel faults is not determined (i.e. barrier or conduit). A tracer test was realized in spring 2022 (47.00, 6.92) on the northern tip of the Neuchâtel fault but none of the monitored springs in the Neuchâtel area showed traces of the tracer. Similarly, a second test was realized further east (47.01, 6.96), near the Monruz Fault without any recovered tracer at the monitored springs (unpublished Master Thesis, University of Neuchâtel, Solène Dal Fitto 2022). This lack of tracer at the different outlets of the major aquifer of the study area emphasises the need to better understand the hydrogeologic functioning of this reservoir. In addition, continuous monitoring of the springs and detailed analysis of their hydrographs would help classify and compare the different hydrotectonic compartments with regards to their hydrogeological properties (i.e. fracturation, karstification, permeability, storativity) (see for example Kovács et al. (2005), by analyzing the response of the karstic spring to a specific precipitation event.

The major outcome of such investigation could be an updated hydrogeological map of the Neuchâtel area, highlighting the different hydrotectonic compartment and their hydrogeological characteristics.

Structural analysis

One aspect that was not covered in this study is a detailed analysis of the deformation structures of the area and how they relate to the observed seismicity. An updated tectonic map of the

area integrating interpretations of various authors (i.e., Sommaruga (1997); Rey and Mosar (2015); Vouillamoz et al. (2016); Gruber (2017); Rime et al. (2019)) and synthesizing findings of the present study could be realized. However, drawing conclusions from an integrated map of this kind may be regarded as questionable due to the low quality of earthquakes (i.e., high uncertainty in their location), and interpretations of the results should be approached with caution. Nevertheless, it would allow to obtain a clearer picture of potentially active structures and their role in accommodating the on-going alpine deformation.

Implication for geothermal energy exploitation

The conclusions of the present study, especially the one regarding the seismic distribution and the sensitivity of the area to elevated hydraulic head variations can be of interest for any future project of geothermal exploitation in the Neuchâtel area. The seismic monitoring has revealed a higher seismic activity below the lake, near the La Lance fault and the Lessy fault. Following suggestions made by Barton (1973), these faults are most likely hydraulically conductive, meaning they favor groundwater circulation. Drilling a deep borehole within such a structure, could result in a relatively high productive well (i.e., high flow rates) without necessitating hydraulic stimulation to enhance reservoir permeability. The downside is that an active fault zone is more likely to be sensitive to external stress changes and could eventually generate more earthquakes to accommodate the stress reorganization. Considering this drawback, a deep borehole could be imagined outside a fault zone, but where hydraulic stimulation may be necessary to enhance the reservoir permeability, which would on one side favor groundwater circulation, but also change the stress state (increasing pore pressure) and potentially induce seismicity. Regarding the sensitivity of the different fault zones to a varying hydraulic head, the La Lance fault, although it is the most active zone, seems to be the less sensitive to a fluctuating groundwater level, and could be thus the most appropriate place to drill considering a pore pressure variation during geothermal exploitation staying within the order of the MPa.

In view of the results of this study, showing the value that can be obtained by carrying out high-resolution seismic monitoring and coupling these results with hydrogeological data, it is advisable that a subsurface exploration strategy for Switzerland includes a densification of the Swiss seismometer network and long-term monitoring in order to lower the earthquake detection limit and improve the accuracy of their localization. It would also be useful to develop the hydrogeological and piezometric monitoring network in Switzerland. This will improve our understanding of tectonically active zones and their activation when pressure variations occur in the subsurface. As demonstrated in this thesis, it will also enable us to estimate the stress conditions at depth, to better anticipate fault stability and to assess their role in terms of fluid circulation at depth. All these elements are important for planning the exploitation of the subsurface.

Appendix A

Development of equation 2.13

Consider the 1D diffusion equation :

$$\frac{\partial p}{\partial t} = D \frac{\partial^2 p}{\partial z^2} \quad (\text{A.1})$$

on the interval $z \in [0, \infty]$ with initial and boundary conditions

$$p(z, 0) = 0, \quad p(\infty, t) = 0, \quad p(0, t) = I(t) \quad (\text{A.2})$$

where p is the pore pressure, z the distance, t the time and $I(t)$ the input signal:

$$I(t) = (1 - H(t - \tau)) P_a(1 - \frac{t}{\tau}) = H(\tau - t) \frac{P_a}{\tau}(\tau - t) \quad (\text{A.3})$$

where $H(u)$ is the Heaviside step function ($H(u) = 1$ when $u > 0$ and $H(u) = 0$ when $u < 0$), τ the duration of the input signal, and P_a the pressure signal at time $t = 0$. Schematically, [A.3](#) is illustrated on Fig. [A.1a](#).

Considering the impulse response function for a pressure diffusion process:

$$p_\delta(z, t) = \frac{1}{2\sqrt{\pi}} \frac{z}{t} \frac{\exp\left(\frac{-z^2}{4Dt}\right)}{\sqrt{Dt}}, \quad (\text{A.4})$$

satisfying:

$$\frac{\partial p_\delta}{\partial t} = D \frac{\partial^2 p_\delta}{\partial z^2} \quad (\text{A.5})$$

with initial and boundary conditions

$$p_\delta(z, t) = 0, \quad p_\delta(\infty, t) = 0, \quad p_\delta(0, t) = \delta(t) \quad (\text{A.6})$$

where $\delta(t)$ is the Dirac function, the solution is obtain by the following convolution:

$$p(z, t) = \int_0^t p_\delta(z, u) I(t - u) du = \int_0^t p_\delta(z, t - u) I(u) du. \quad (\text{A.7})$$

Substituting [A.3](#) into [A.7](#) results in the superposition of the following two classical solutions : [A.8](#) (constant step, $I_1(t) = P_a$) and [A.9](#) (linear input, $I_2(t) = mt$) ([Crank, 1975](#)) (Fig. [A.1d](#)).

$$p_1(z, t) = P_a \operatorname{erfc} \left(\frac{z}{2\sqrt{Dt}} \right) \quad (\text{A.8})$$

$$p_2(z, t) = mt \left\{ \left(1 + \frac{z^2}{2Dt} \right) \operatorname{erfc} \left(\frac{z}{2\sqrt{Dt}} \right) - \frac{z}{\sqrt{\pi Dt}} \exp \left(\frac{-z^2}{4Dt} \right) \right\} \quad (\text{A.9})$$

Subtracting [A.9](#) from [A.8](#) corresponds to the response of a pressure step decrease linearly in time. The response of the input signal $H(t - \tau)$ $I_2(t - \tau)$ (Fig. [A.1e](#)) is added to the previous subtraction to suppress the constant linear decrease from the time $t = \tau$, when $P_a = 0$. The final solution for the response to a pressure input signal of the form $I(t)$ reads as follows:

$$p(z, t) = p_1(z, t) - p_2(z, t) + H(t - \tau)p_2(z, t - \tau) \quad (\text{A.10})$$

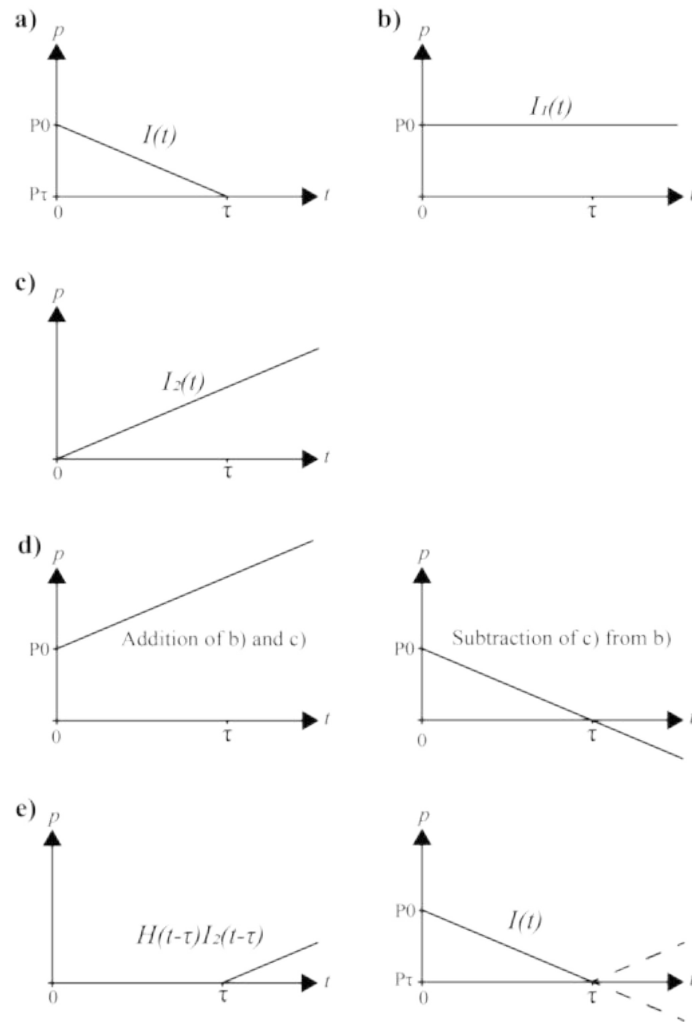


Figure A.1: Graphic representation of the different input signals (a) $I(t)$, b) $I_1(t)$, c) $I_2(t)$) and illustration of the superposition principle (d)) and how it leads to the wanted input signal $I(t)$ (e)).

Appendix B

Tracer Tests

B.1 Gouffre du Pertuis Injections

Two tracer tests were realized in the vicinity of the Ferrière Fault. Hydrogeologic monitoring during the tracer tests included a close observation of the recovery of the tracer at the springs with field fluorometer. During these tests, additional observation points were monitored and water samples were collected. Data from the tracer test comprises measurements from the field fluorometer as well as punctual water samples. The water samples are analyzed in a laboratory spectrofluorometer, the measured values are then used to calibrate the data from the field fluorometer. The concentration C measured with the spectrofluorometer, along with the average spring flowrate Q during the period of restitution (time between t_0 and t_i) is used to calculate the recovered mass using the following equation:

$$M_r = \int_{t_0}^{t_i} C \cdot Q \, dt \quad (\text{B.1})$$

The recovery rate R , usually presented as a percentage, is the ratio of the recovered mass M_r over the injected mass M_i multiplied by 100 :

$$R = \frac{M_r}{M_i} \cdot 100 \quad (\text{B.2})$$

Tab. [B.1](#) lists the different tracer tests, the injection point, the observation point, the tracer injected and whether the tracer was detected or not as well as the recovery rate.

Fig. [B.1](#) shows the recovery curve for the two injections mentioned in Tab. [B.1](#) and the calculation of the recovery rate using Eqs. [\(B.1\)\(B.2\)](#)

Table B.1: Location of injection and observation points and results of the tracer tests

	Lat.	Long.	Alt.	Tracer	Time	
Injection 1	47.09423	6.9358	1016	2 kg Uranine	22.03.2021 13:30	
Observation points				Detection	First arrival	Recovered mass
Serrière	46.9856	6.9013	466.8	Yes	26.03.2021 10:00	17 %
Seyon	46.9896	6.921	437	No		
Ecluse	46.9926	6.9248	446	No		
Merdasson	46.9703	6.8232	577.2	No		
Raissette	47.1739	7.0524	714.4	No		
Puits des Sauges	47.1386	6.9758	795.4	No		
Injection 2	47.09423	6.9358	1016	3 kg Uranine	30.04.2021 14:00	
Observation points				Detection	First arrival	Recovered mass
Serrière	46.9856	6.9013	466.8	Yes	04.05.2021 08:00	40 %
Seyon	46.9896	6.921	437	No		
Ecluse	46.9926	6.9248	446	No		
Merdasson	46.9703	6.8232	577.2	No		
Pockmark Serrière	46.9727	6.8959	340	No		

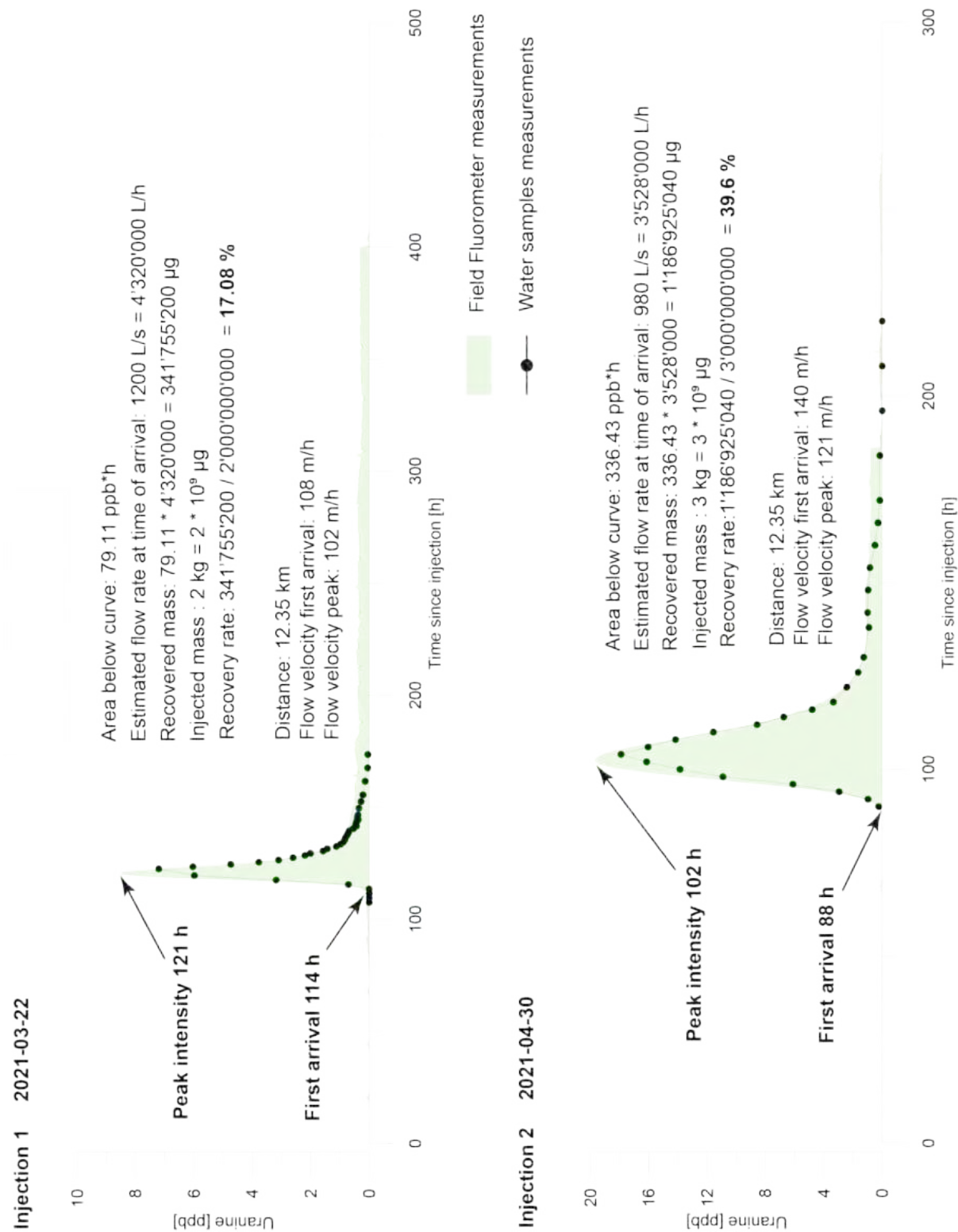


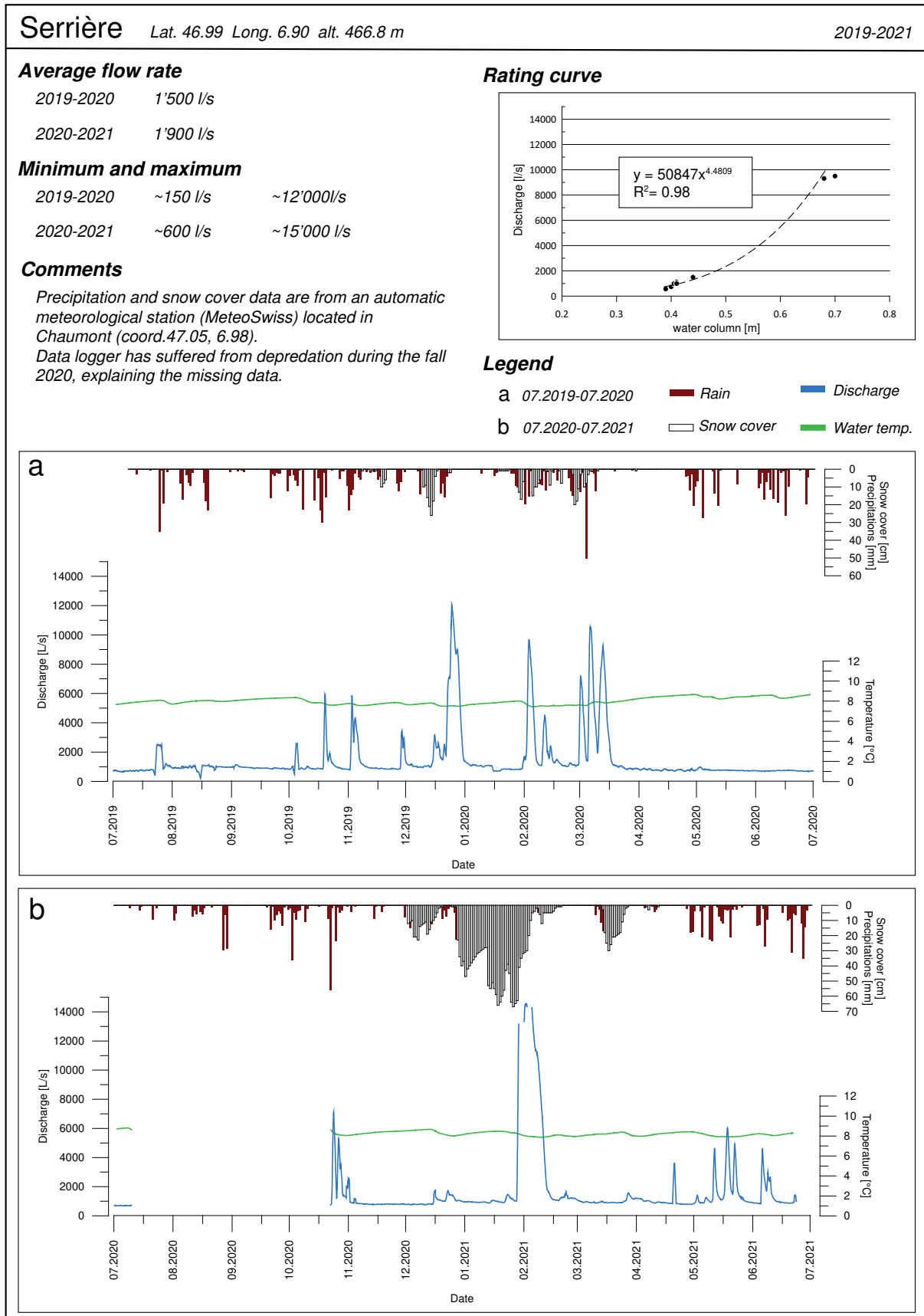
Figure B.1: Recovery curves of injections 1 and 2 and calculated flow velocities and recovery rates.

Appendix C

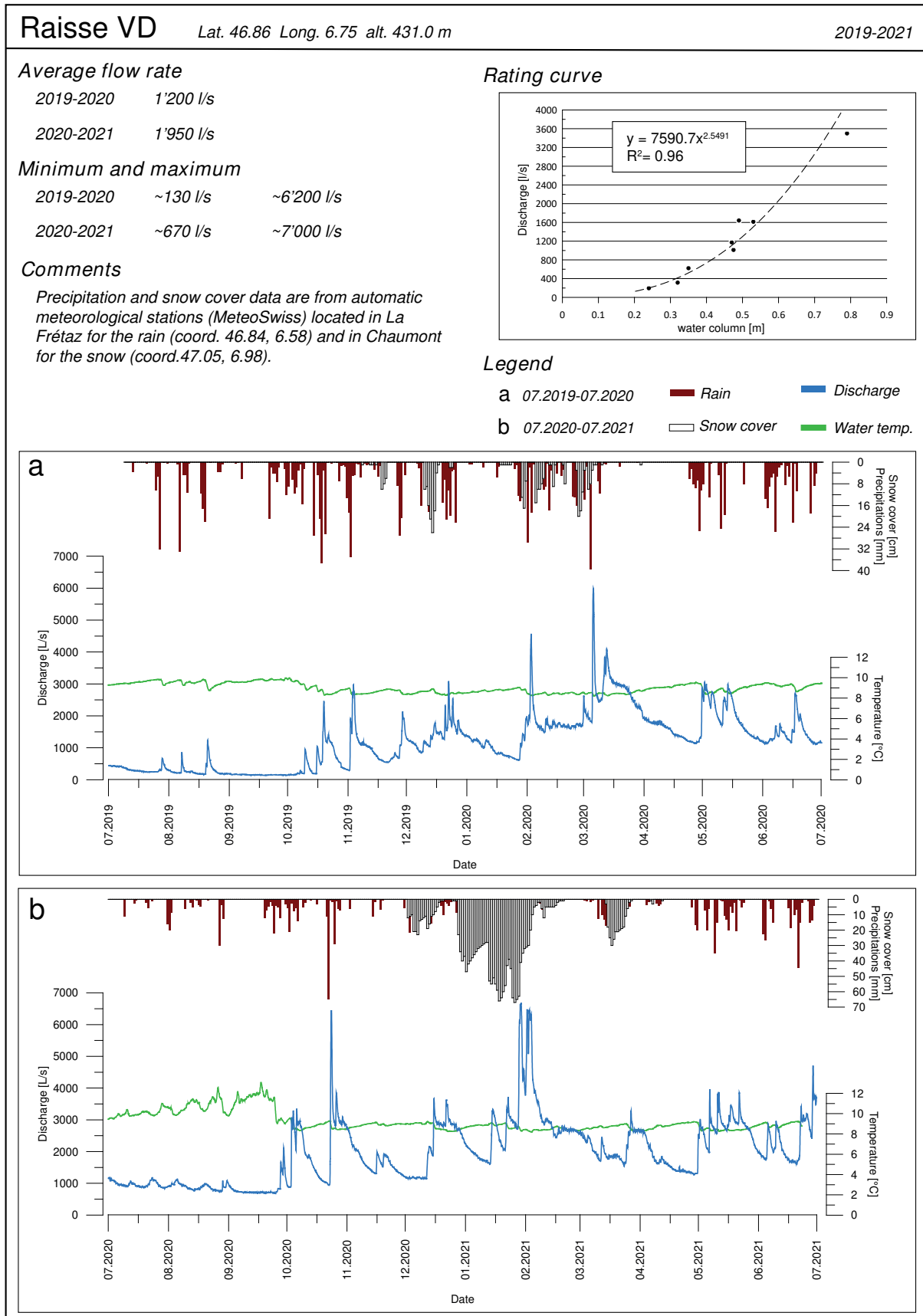
Hydrographs

A detailed analysis of the springs of the Serrière (C.1), the Raisse VD (C.2), the Dia (C.3) and the Torrent (C.4) is presented in Chapter 4. Other springs are briefly introduced here. Similarly to the springs mentioned in Chapter 4, the Raisse Fleurier spring (C.5) located in the Malm, shows rapid and important flow rate variations and represents a typical karst hydrograph. The Tannes spring (C.6) is in Cretaceous limestones, underlying Quaternary deposits. The hydrograph represents the water level within the catchment chamber. During the months of spring and summer, the water level is not affected by precipitations. The water level increases rapidly as a response to the infiltration of the melted snow. The hypothesis regarding the Tannes spring is that the Quaternary sediments act as a delaying factor regarding infiltration from the precipitation into the Cretaceous sediments resulting in a quasi-constant water level. The rapid level variations can be explained by a possible connection to the Malm aquifer, underlying the Cretaceous, when the hydraulic gradient within the Malm increases enough to reach the Cretaceous aquifer. The Bonvillars (C.7) and the Champagne (C.8) springs are located near the limit between the Quaternary and Cretaceous deposits. These springs have been monitored to observe whether they are also connected to the underlying Malm aquifer. For Bonvillars, the fluctuation of the water level is moderated and suggests a similar situation as in Les Tannes, however, no important flooding occurred during the monitoring period to verify the reaction of the Bonvillars spring in comparison to the Tannes spring. For Champagne, the hydrograph shows important variations, suggesting a clear karstic origin, most certainly from the Cretaceous.

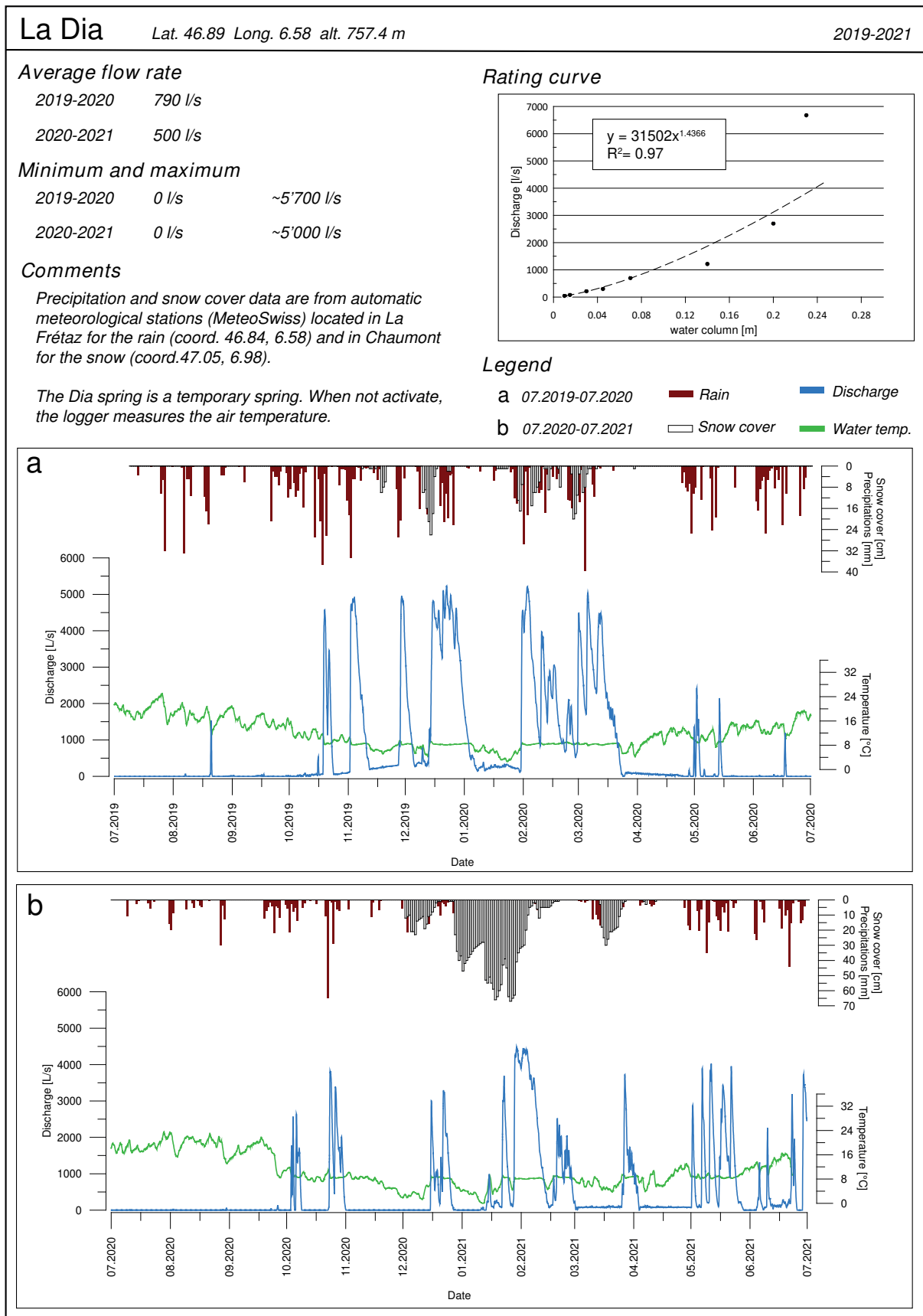
C.1 Serrière



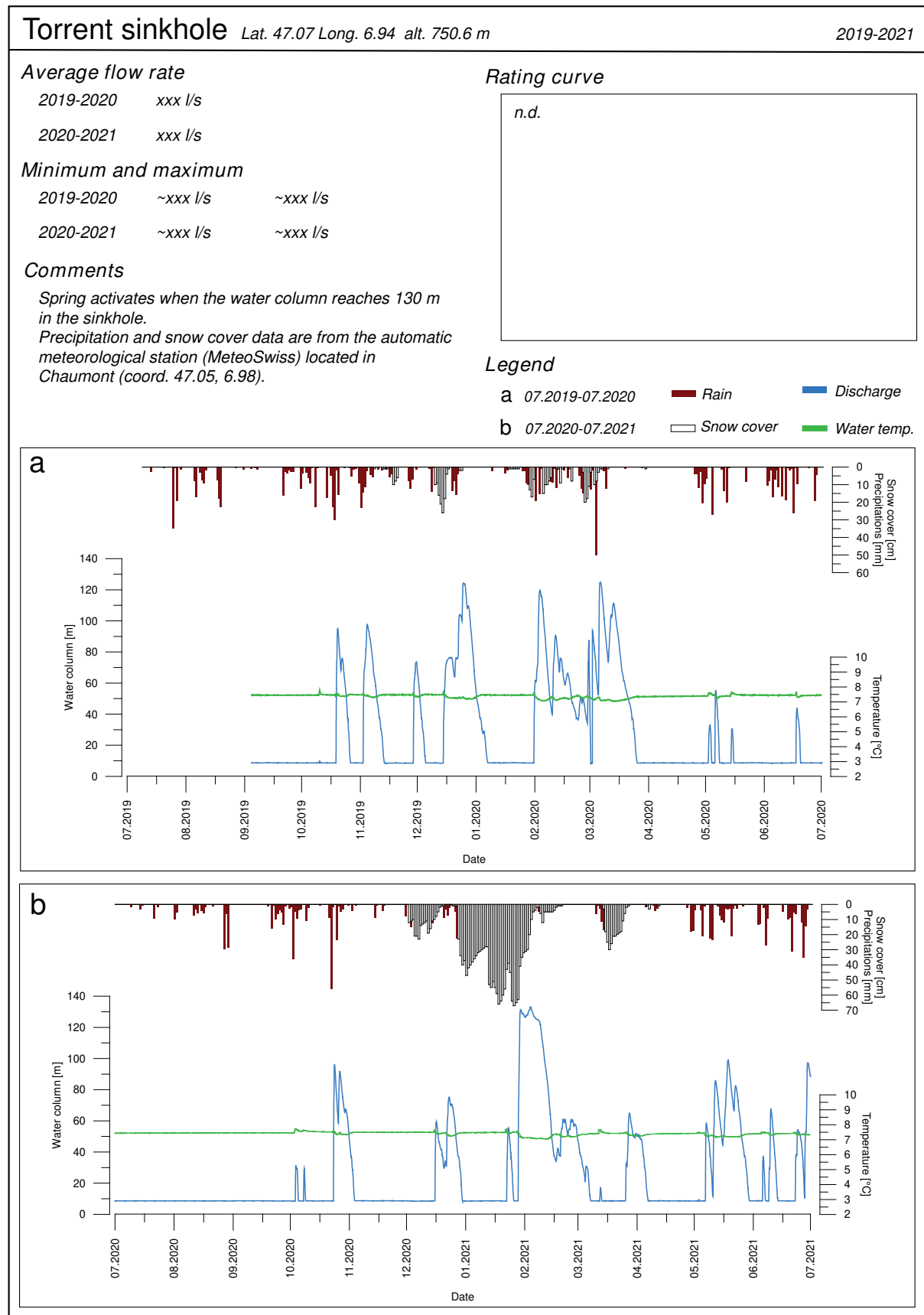
C.2 Raisse VD



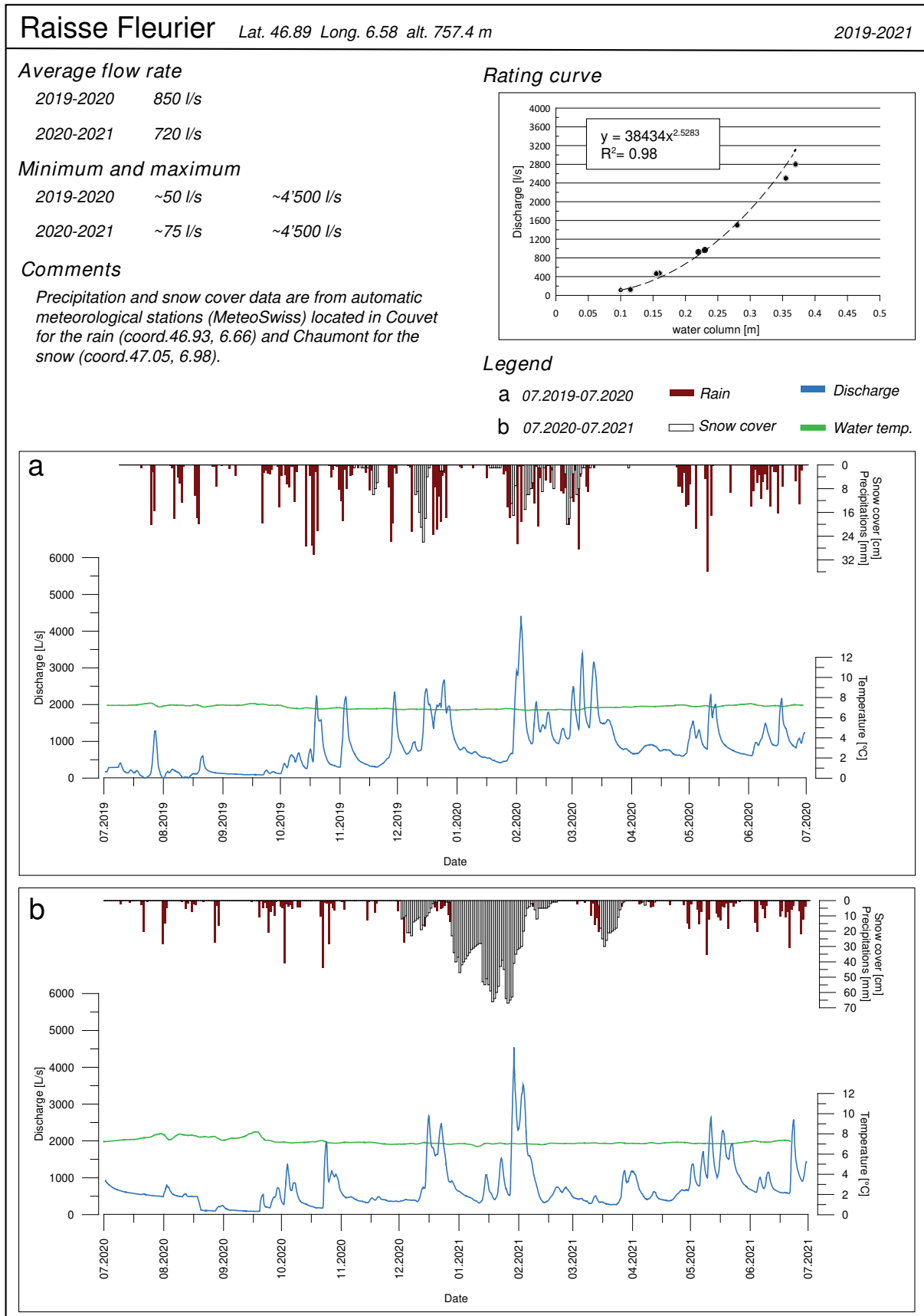
C.3 Dia



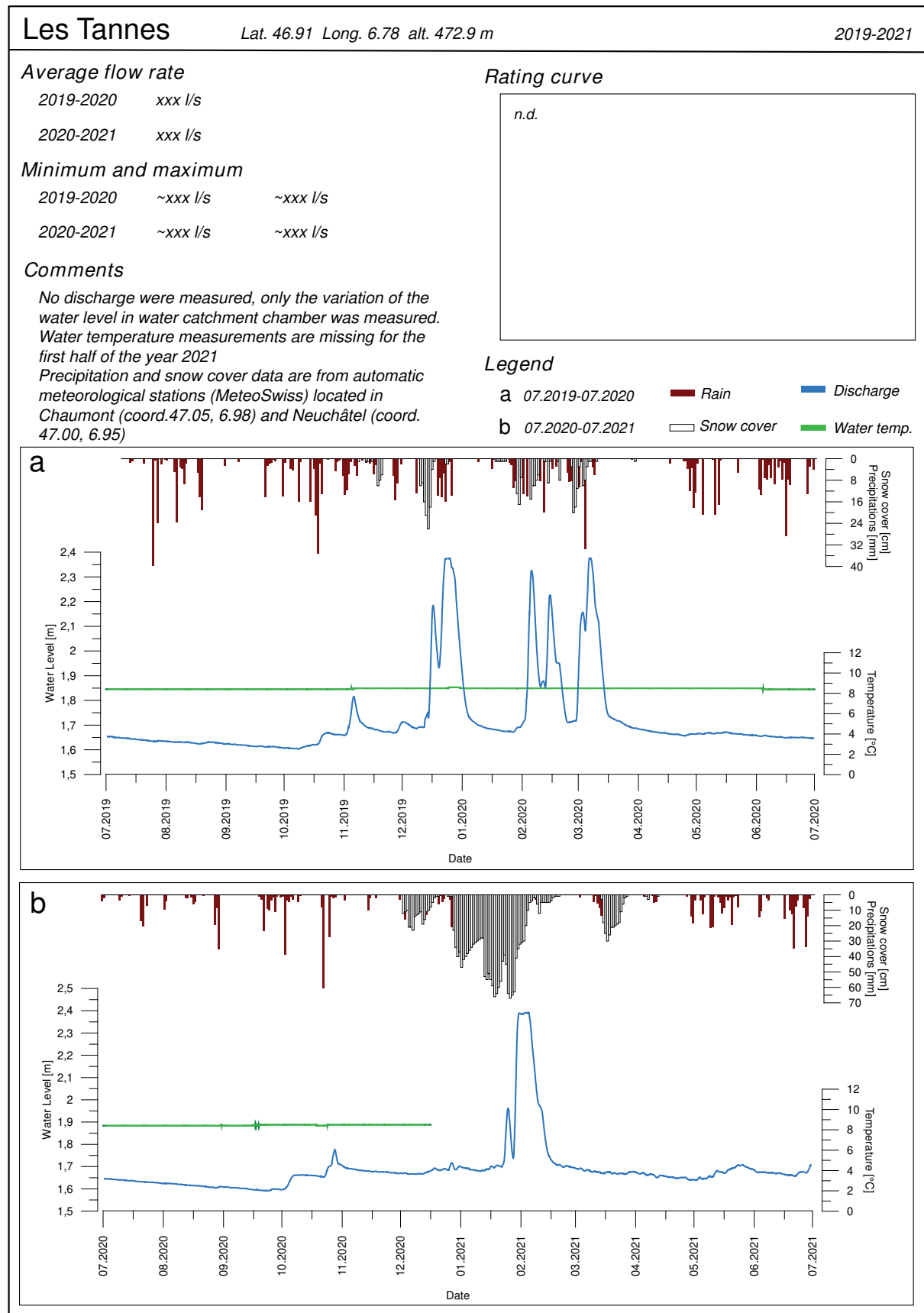
C.4 Torrent



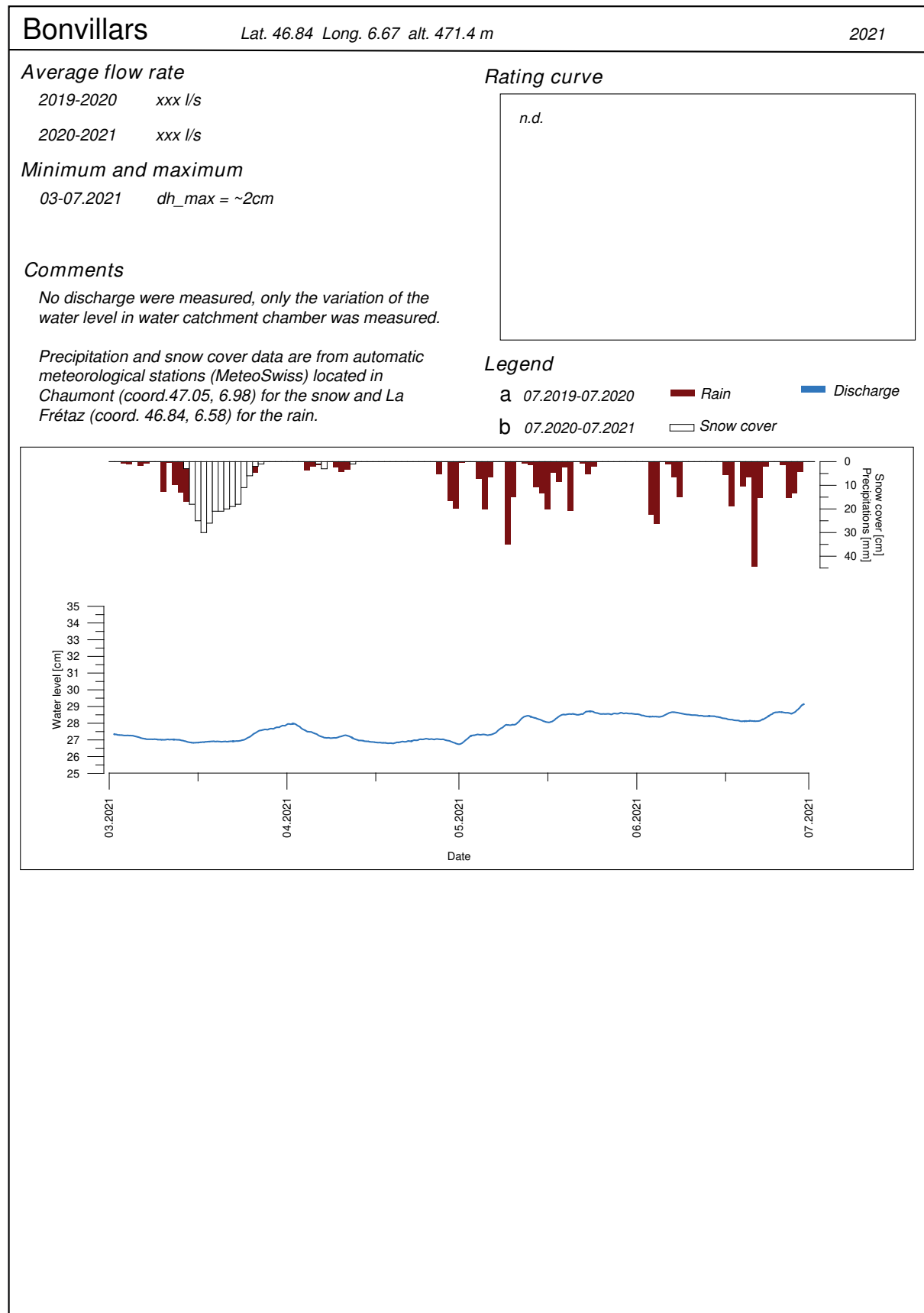
C.5 Raisse Fleurier



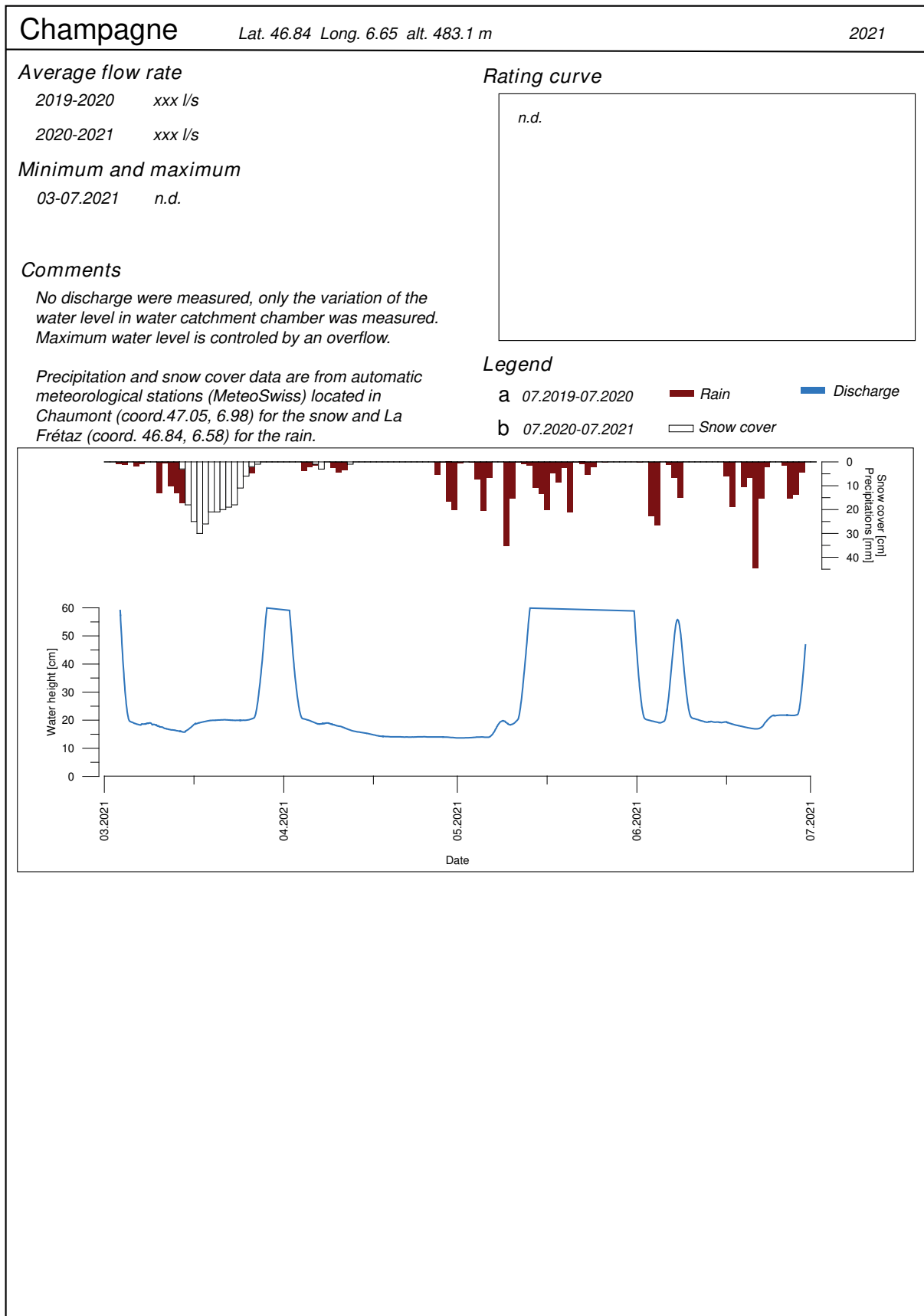
C.6 Tannes



C.7 Bonvillars



C.8 Champagne



Appendix D

STA/LTA

Different parameters used in this study for STA/LTA coincidence triggering (D.1) and cross-correlation (G).

D.1 STA/LTA parameters

STA/LTA trigger algorithm has been used through the ObsPy toolbox (Beyreuther et al., 2010).

Table D.1: STA/LTA parameters

Channel	Z
Bandpass filter (Hz)	1-30
Minimum stations	3
STA window (s)	0.5
LTA window (s)	15
Trigger "ON"	3.5
Trigger "OFF"	2.0

Appendix E

List of detected earthquakes

Earthquake catalogue for the Neuchâtel area from July 2019 to July 2021. Depth is in km below sea level. LQ: event identified as a local earthquake, LU: event identified as local but uncertain about the origin. Clusters are attributed after performing the inter-event cross-correlation. AOI means area of interest and corresponds to the sub-region the earthquake occurred in. The two last columns are water level variations in the Torrent sinkhole in meters, at the time of the earthquake (dh) and the maximum variation over the last 24 hours (Max dh 24h).

ID	Date and time (UTC)	Latitude	Longitude	Depth	ML	Type	Cluster	Tribe	AOI	dh	Max dh 24h
1	2019-07-29 19:24:01	46.821	6.865	2.2	0.7	LQ				0	0
2	2019-08-02 20:45:37	46.843	6.69	1.2	0.8	LQ	12	LAN	LAN	0	0
3	2019-09-12 21:36:22	46.983	6.946	0	-0.4	LU	3	FER	NEU	0	0
4	2019-10-03 16:31:26	47.126	6.818	8.1	2.1	LQ				0	0
5	2019-10-06 05:54:54	46.842	6.691	1.4	-0.5	LQ	12	LAN	LAN	0	0
6	2019-10-06 07:20:52	46.849	6.778	8	0.8	LQ	15	LAN	LAN	0	0
7	2019-10-08 18:34:33	46.856	6.711	0.1	0.3	LU	12	LAN	LAN	0	0
8	2019-10-14 04:36:15	46.84	6.704	0.7	0.1	LQ	12	LAN	LAN	0	0
9	2019-10-21 02:44:28	46.921	6.805	0	-0.4	LU			LESSY	68.99	85.88
10	2019-10-23 23:50:00	46.874	6.821	0	0	LQ				49.44	62.11
11	2019-11-18 23:10:33	46.855	6.815	0	0	LQ			LAN	0	0
12	2019-11-25 11:23:55	46.846	6.747	11.8	0.6	LQ	15	LAN	LAN	0	0
13	2019-11-25 12:43:34	46.863	6.809	2.2	0.1	LQ			LAN	0	0
14	2020-01-27 02:18:49	46.85	6.746	6.7	0.9	LQ	6	LAN	LAN	0	0
15	2020-02-05 11:46:35	46.864	6.713	9.5	1.2	LQ	6	LAN	LAN	94.35	107.25
16	2020-02-05 23:07:04	46.849	6.825	8	0.6	LU			LAN	86.83	103.16
17	2020-02-06 23:20:41	46.805	6.719	8	0.2	LQ				71.53	86.83
18	2020-02-14 02:55:34	46.862	6.731	8.6	1.1	LQ	6	LAN	LAN	61.89	69.14
19	2020-02-15 05:00:45	46.788	6.685	1.4	0	LU	2	LAN		64.97	67.25
20	2020-02-15 05:01:49	46.789	6.676	2.2	0.4	LQ	2	LAN		64.97	67.25
21	2020-02-27 00:05:03	46.803	6.741	8.1	0.1	LU				38.15	41.44
22	2020-03-07 20:07:29	46.864	6.709	0.3	0.4	LQ	13	LAN	LAN	103.24	115.11
23	2020-04-06 00:39:13	46.834	6.826	0.1	-0.1	LQ			LAN	0	0
24	2020-04-06 02:13:03	46.916	6.933	2.3	0.3	LQ				0	0
25	2020-05-20 17:27:47	46.875	6.855	0	0.5	LQ	11			0	0
26	2020-06-05 15:14:03	46.855	6.73	2	0.4	LQ	15	LAN	LAN	0	0
27	2020-06-22 18:13:02	46.792	6.723	0.5	0.1	LQ				0	0
28	2020-07-04 17:58:49	46.889	6.815	1.5	0.1	LQ	14	BOU	LESSY	0	0
29	2020-07-06 12:00:37	46.893	6.811	1.5	0.4	LU	14	BOU	LESSY	0	0
30	2020-07-07 11:03:35	46.836	6.901	3.1	0.6	LU	4		LAN	0	0
31	2020-07-08 06:33:24	46.893	6.81	1.7	-0.3	LU	14	BOU	LESSY	0	0
32	2020-07-11 05:26:00	47.061	6.866	0.9	0.5	LQ	5	FER	FER	0	0

33	2020-07-12 06:12:07	46.879	6.86	0	0.8	LQ	11	EST		0	0	0
34	2020-07-12 12:31:25	47.083	6.878	2.2	2.5	LQ	5	FER	FER	0	0	0
35	2020-07-12 13:13:28	47.063	6.862	9.2	0.9	LQ	5	FER	FER	0	0	0
36	2020-07-12 13:43:36	47.068	6.872	4.3	0.5	LQ	5	FER	FER	0	0	0
37	2020-07-12 19:16:32	46.876	6.585	31.9	0.4	LQ				0	0	0
38	2020-07-15 03:01:05	47.068	6.871	0.5	1.1	LQ	5		FER	0	0	0
39	2020-07-16 20:34:08	46.898	6.773	1	-1	LQ			LESSY	0	0	0
40	2020-07-18 20:22:06	46.766	6.793	3.1	0.4	LQ				0	0	0
41	2020-07-26 00:52:47	46.883	6.884	1.8	-0.5	LQ	11			0	0	0
42	2020-08-13 19:13:16	46.9	6.72	0	-0.7	LU			LESSY	0	0	0
43	2020-08-16 01:12:21	47.112	6.349	20	0.3	LU				0	0	0
44	2020-08-19 22:15:34	46.867	6.715	1.1	-0.7	LQ			LAN	0	0	0
45	2020-09-04 18:05:37	46.847	6.792	12.3	-0.2	LU			LAN	0	0	0
46	2020-09-05 03:42:43	46.735	6.754	0.6	0.5	LQ				0	0	0
47	2020-09-05 20:01:38	47.116	6.706	12.7	1	LU				0	0	0
48	2020-09-28 06:19:19	46.834	6.884	1.8	0.4	LQ	4		LAN	0	0	0
49	2020-09-28 06:22:57	46.837	6.894	2.7	0.9	LQ	4		LAN	0	0	0
50	2020-09-28 06:27:26	46.835	6.889	2.8	1.3	LQ	4		LAN	0	0	0
51	2020-10-08 22:25:23	46.914	6.805	0.8	-0.1	LQ			LESSY	1.61	20.71	
52	2020-10-13 16:09:46	46.826	6.762	0.7	0.1	LQ			LAN	0	0	0
53	2020-10-15 00:11:50	46.836	6.867	1.3	0.6	LQ	4		LAN	0	0	0
54	2020-10-15 00:16:56	46.833	6.877	1.9	1.1	LQ	4		LAN	0	0	0
55	2020-10-15 00:18:37	46.831	6.874	0.7	0.7	LQ	4		LAN	0	0	0
56	2020-10-15 00:22:00	46.838	6.87	2.7	0.9	LQ	4		LAN	0	0	0
57	2020-10-15 00:59:02	46.833	6.882	1.4	1.1	LQ	4		LAN	0	0	0
58	2020-10-15 01:01:57	46.84	6.891	4	0.9	LQ	4		LAN	0	0	0
59	2020-10-18 03:09:18	46.783	6.72	3.9	-0.6	LQ	2	LAN		0	0	0
60	2020-10-20 00:40:49	46.869	6.814	1.4	-0.8	LU				0	0	0
61	2020-10-20 17:52:31	46.832	6.895	2.6	1.1	LQ	4		LAN	0	0	0
62	2020-10-20 18:06:18	46.832	6.89	2.7	1.8	LQ	4	EST	LAN	0	0	0
63	2020-10-20 18:36:42	46.833	6.896	2.6	0.8	LQ	4		LAN	0	0	0
64	2020-10-20 18:43:10	46.823	6.888	2.1	0.3	LU	4		LAN	0	0	0
65	2020-11-05 23:28:56	46.794	6.676	0	-0.1	LQ				0	0	0

66	2020-11-10 10:00:50	46.939	6.537	0.9	1.5	LQ		0	0
67	2020-11-22 05:02:04	46.883	6.795	0	-0.7	LU		0	0
68	2020-11-23 00:48:30	46.875	6.791	0.4	-0.7	LU		0	0
69	2020-12-05 23:23:28	46.792	7.066	12	0.6	LQ		0	0
70	2020-12-06 09:36:12	46.98	6.877	0.1	0.5	LQ		0	0
71	2020-12-14 22:09:25	46.714	6.657	2.1	1.6	LQ	EST	0	0
72	2020-12-15 01:41:04	46.743	6.698	0	0.3	LQ	7	0	0
73	2021-01-09 04:03:48	46.808	6.714	3.3	0.6	LQ		0	0
74	2021-01-18 18:31:23	46.809	6.715	1.7	-0.1	LQ		0	0
75	2021-01-24 01:29:33	47.087	6.853	11	0	LQ	FER	40.89	46.67
76	2021-01-27 21:59:33	46.799	6.745	2.2	0.7	LQ		0	0
77	2021-02-05 14:12:09	46.862	6.717	0	0.7	LQ	13 LAN	119.29	122.77
78	2021-02-06 02:44:09	46.867	6.707	0	0.8	LQ	13 LAN	118	120.73
79	2021-02-07 09:37:20	47.081	7.09	2.8	2.8	LQ	17 FER	116.33	117.51
80	2021-02-10 16:05:09	47.019	7.004	0.6	0.6	LQ	9 FER	99.14	108.71
81	2021-02-11 05:39:51	47.018	7.056	2.1	0.9	LQ	9 FER	92.61	102.77
82	2021-02-11 22:22:00	46.99	7.179	8	1	LQ	8	86.03	96.62
83	2021-02-11 22:45:42	47.002	7.188	8.1	0.5	LU	8	85.78	96.13
84	2021-02-13 11:14:21	46.991	7.016	2.2	-0.1	LU	9 FER	69.91	81.06
85	2021-02-16 03:23:12	47.032	7.065	2.4	0.5	LQ	SBL	31.31	42.58
86	2021-02-22 02:37:35	46.957	6.914	0	-0.6	LQ		52.13	52.4
87	2021-02-24 18:07:16	46.714	6.68	2.2	1	LQ	7	45.69	46.08
88	2021-03-07 15:31:58	46.845	6.736	4.1	0.2	LQ	1 LAN	0	12.18
89	2021-03-07 15:41:16	46.845	6.738	3.3	0.6	LQ	1 LAN	0	12.18
90	2021-03-07 18:43:01	46.834	6.753	2.7	0.2	LQ	LAN	0	9.7
91	2021-03-14 02:45:38	46.845	6.737	2.9	1.2	LQ	1 LAN	0	4.64
92	2021-03-14 05:17:33	46.853	6.721	3.5	1.6	LQ	1 LAN	0	3.33
93	2021-03-14 05:32:46	46.84	6.746	1.5	0.5	LQ	1 LAN	0	2.98
94	2021-03-17 00:09:27	47.07	7.092	1.3	0.3	LQ	NEUVE	0	0
95	2021-03-18 05:07:38	46.85	6.729	2.2	0.6	LQ	1 LAN	0	0
96	2021-03-25 12:49:41	46.846	6.729	3.4	1.3	LQ	1 LAN	0	0
97	2021-03-31 22:52:44	46.848	6.775	0	0.4	LQ	LAN	41.88	43.01
98	2021-04-01 14:36:17	47.025	6.707	2.2	1	LU		41.37	42.71

99	2021-04-29 05:28:02	47.181	6.939	8	1.1	LQ			0	0
100	2021-05-02 09:45:57	46.859	6.717	1.8	1.1	LQ	LAN		0	0
101	2021-05-21 01:52:55	46.922	6.819	0	2.4	LQ	BOU	16	65.08	76.5
102	2021-05-22 15:05:27	46.925	6.842	0	0.8	LQ	BOU	16	73.62	73.55
103	2021-06-05 17:25:36	46.843	6.733	0.8	1.7	LQ	LAN	1	0	0
104	2021-06-06 13:21:58	46.79	6.673	1.3	0.7	LQ	LAN	2	22.24	28.35
105	2021-06-07 16:11:22	46.752	6.617	2.1	1.7	LQ	LAN, EST	10	0	20.06
106	2021-06-09 08:30:07	47.02	6.937	0	1.9	LQ	FER	3	18.88	17.99
107	2021-06-11 13:08:04	46.991	6.93	0	0.3	LQ	FER	3	40.86	56.17
108	2021-06-12 22:40:04	46.997	6.925	0.6	0.8	LQ	FER	3	17.07	34.66
109	2021-06-15 14:18:05	46.99	6.927	0	0.9	LQ	FER	3	0	0
110	2021-06-15 15:21:58	46.991	6.926	0	0.6	LQ	FER	3	0	0
111	2021-06-15 16:37:55	46.918	6.797	0	0.1	LQ	LESSY	0	0	0
112	2021-06-15 23:04:01	46.993	6.926	0	0.7	LQ	FER	3	0	0
113	2021-06-27 10:55:08	47.015	6.931	0	1.9	LQ	FER	3	14.86	34.95
114	2021-06-27 17:12:47	46.993	6.924	0	0.5	LQ	FER	3	6.02	30.78
115	2021-06-28 04:52:09	46.92	6.823	0	0.6	LQ	LESSY	16	35.52	34.24
116	2021-06-28 13:18:32	46.959	6.858	14.6	1.5	LQ	LESSY	16	44.18	44.13
117	2021-06-28 18:55:21	46.891	6.792	10.9	1.4	LQ	LESSY		47.58	46.51
118	2021-06-30 23:49:02	46.736	6.689	2.1	0.1	LQ	LAN	10	79.79	87.27
119	2021-06-30 23:50:31	47.069	6.87	12	1	LQ	FER	5	79.79	87.27
120	2021-07-03 17:03:31	46.903	6.868	0	0.1	LQ			51.81	65.27
121	2021-07-04 03:49:56	46.687	6.662	10.2	0.1	LQ			45.84	57.62
122	2021-07-05 14:18:57	46.802	6.809	5.1	0.5	LQ			28.54	38.76
123	2021-07-13 11:15:57	46.982	6.925	0.1	1.3	LQ	FER	3	101.27	99.42
124	2021-07-14 01:38:01	46.897	6.885	2.3	0.3	LQ			116.02	116.38
125	2021-07-17 22:45:25	46.735	6.672	2.1	0.4	LQ	LAN	10	115.8	118.15
126	2021-07-21 03:01:40	47.069	7.071	2.1	1	LQ	FER	17	84.2	94.58
127	2021-07-22 15:02:45	46.866	6.825	0	0.5	LQ			66.64	79.19
128	2021-07-24 15:26:05	46.893	6.913	9	1.3	LQ			35.39	51.61
129	2021-07-27 14:11:06	46.796	6.826	2.3	0.8	LQ			15.52	21.54

Appendix F

Quality ranking

Table showing the values of parameters to estimate the quality of the earthquake location for each detected event and its score according to the threshold value defined in Table 3.4, the score is 1 if it is equal or higher and 0 if it is lower. A score of 0.5 is attributed to parameters 1 and 2 (Number of station and Number of P and S phases) if the value is just below (i.e. 5 and 7, respectively). The sum of each individual score gives the final score and attributes a quality to each event. 1-2: Low quality, 3-4: Medium quality and 5: High quality

ID	Num. stations	Num. P and S phases	Dist. closest station	Az. Gap	RMS of sol.	Score num. stations	Score num. phases	Score dist. closest station	score Az. Gap	Score RMS	Final score	Cluster
1	7	13	2	191	0.3	1	1	1	0	0	3	
2	6	11	1	134	0.1	1	1	1	0	1	4	12
3	4	8	2	307	0.2	0	1	0	0	1	2	3
4	12	21	13	296	0.3	1	1	0	0	0	2	
5	5	10	0	149	0.1	0.5	1	1	0	1	3.5	12
6	8	16	5	153	0.1	1	1	1	0	1	4	15
7	3	7	2	300	0	0	0.5	0	0	1	1.5	12
8	7	13	1	218	0.1	1	1	1	0	1	4	12
9	4	7	2	247	0	0	0.5	0	0	1	1.5	
10	11	18	5	148	0.2	1	1	0	0	1	3	
11	11	20	4	148	0.3	1	1	0	0	0	2	
12	10	17	3	167	0.4	1	1	1	0	0	3	15
13	6	8	5	284	0.2	1	1	0	0	1	3	
14	16	27	3	134	0.1	1	1	1	0	1	4	6
15	12	21	0	63	0.4	1	1	1	1	0	4	6
16	4	6	7	330	0	0	0	1	0	1	2	
17	6	9	6	327	0.1	1	1	1	0	1	4	
18	11	22	1	109	0.4	1	1	1	1	0	4	6
19	3	12	6	341	0	0	1	0	0	1	2	2
20	6	11	6	304	0.1	1	1	0	0	1	3	2
21	3	4	8	326	0	0	0	1	0	1	2	
22	9	15	1	71	0.1	1	1	0	1	1	4	13
23	11	20	1	74	0.2	1	1	0	1	1	4	
24	8	12	8	282	0.2	1	1	0	0	1	3	
25	11	16	2	121	0.2	1	1	0	0	1	3	11
26	5	9	1	213	0	0.5	1	1	0	1	3.5	15
27	8	14	4	229	0.2	1	1	0	0	1	3	
28	7	12	4	267	0.1	1	1	0	0	1	3	14
29	5	8	4	251	0	0.5	1	0	0	1	2.5	14
30	4	9	4	259	0.1	0	1	1	0	1	3	4

31	6	10	4	261	0.1	1	1	0	0	1	3	14
32	8	16	5	211	0.1	1	1	0	0	1	3	5
33	19	31	6	124	0.2	1	1	0	0	1	3	11
34	25	50	6	242	0.2	1	1	0	0	1	3	5
35	8	13	5	210	0.3	1	1	1	0	0	3	5
36	7	11	5	223	0.2	1	1	1	0	1	4	5
37	6	7	3	317	0.3	1	0.5	1	0	0	2.5	
38	20	31	5	222	0.2	1	1	0	0	1	3	5
39	4	6	1	195	0	0	0	1	0	1	2	
40	9	14	3	261	0.2	1	1	1	0	1	4	
41	3	8	3	279	0	0	1	0	0	1	2	11
42	5	7	3	131	0.1	0.5	0.5	0	0	1	2	
43	3	6	33	343	0.1	0	0	0	0	1	1	
44	6	10	1	124	0.1	1	1	1	0	1	4	
45	3	6	5	299	0.2	0	0	1	0	1	2	
46	9	16	11	287	0.1	1	1	0	0	1	3	
47	5	8	16	340	0	0.5	1	1	0	1	3.5	
48	7	13	3	241	0.1	1	1	0	0	1	3	4
49	6	11	3	252	0.2	1	1	1	0	1	4	4
50	11	19	3	249	0.1	1	1	1	0	1	4	4
51	3	5	3	260	0	0	0	0	0	1	1	
52	8	12	5	300	0.1	1	1	0	0	1	3	
53	10	13	2	180	0.3	1	1	0	0	0	2	4
54	9	15	3	197	0.1	1	1	0	0	1	3	4
55	8	12	3	220	0.3	1	1	0	0	0	2	4
56	8	14	2	194	0.3	1	1	1	0	0	3	4
57	11	18	3	203	0.2	1	1	0	0	1	3	4
58	6	9	3	259	0.1	1	1	1	0	1	4	4
59	6	11	4	245	0.2	1	1	1	0	1	4	2
60	5	8	4	155	0.1	0.5	1	0	0	1	2.5	
61	11	18	4	272	0.1	1	1	0	0	1	3	4
62	17	26	3	225	0.2	1	1	1	0	1	4	4
63	8	14	4	271	0.1	1	1	0	0	1	3	4

64	3	5	4	320	0	0	0	0	1	1	4
65	9	14	5	246	0.1	1	1	0	1	3	
66	12	17	9	283	0.3	1	1	0	0	2	
67	3	6	4	282	0.1	0	0	0	1	1	
68	3	7	4	290	0	0	0.5	0	1	1.5	
69	11	19	16	302	0.5	1	1	1	0	3	
70	6	9	0	147	0	1	1	1	1	4	
71	12	20	12	295	0.2	1	1	0	1	3	7
72	5	7	11	331	0.2	0.5	0.5	0	1	2	7
73	6	9	4	209	0.1	1	1	1	1	4	
74	5	7	4	190	0.1	0.5	0.5	0	1	2	
75	7	12	7	303	0.1	1	1	1	1	4	
76	8	11	6	298	0.2	1	1	0	1	3	
77	6	12	0	160	0.1	1	1	1	1	4	13
78	5	9	1	149	0.1	0.5	1	0	1	2.5	13
79	12	17	14	336	0.3	1	1	0	0	2	17
80	4	6	6	334	0.2	0	0	0	1	1	9
81	4	7	9	345	0.1	0	0.5	0	1	1.5	9
82	4	7	19	355	0.2	0	0.5	0	1	1.5	8
83	4	6	18	354	0.2	0	0	0	1	1	8
84	4	7	5	350	0.4	0	0.5	0	0	0.5	9
85	3	6	10	348	0.1	0	0	0	1	1	
86	5	9	3	291	0.2	0.5	1	0	1	2.5	
87	6	11	11	313	0.1	1	1	0	1	3	7
88	5	10	3	250	0.1	0.5	1	1	1	3.5	1
89	7	12	3	140	0.1	1	1	1	1	4	1
90	6	9	4	286	0.1	1	1	1	1	4	
91	8	15	3	141	0.1	1	1	1	1	4	1
92	9	13	1	110	0.1	1	1	1	1	5	1
93	6	11	3	165	0.1	1	1	0	1	3	1
94	5	9	13	338	0.1	0.5	1	0	1	2.5	
95	10	18	2	123	0.2	1	1	1	1	4	1
96	9	14	3	101	0.1	1	1	1	1	5	1

97	7	11	4	268	0.1	1	1	0	0	1	3	
98	3	6	10	346	0.2	0	0	0	0	1	1	
99	8	16	29	345	0.1	1	1	0	0	1	3	
100	15	25	0	68	0.2	1	1	1	1	1	5	
101	19	34	2	103	0.3	1	1	0	1	0	3	16
102	8	14	3	297	0.2	1	1	0	0	1	3	16
103	18	29	3	108	0.2	1	1	0	1	1	4	1
104	8	12	6	252	0.1	1	1	0	0	1	3	2
105	13	18	11	290	0.2	1	1	0	0	1	3	10
106	19	37	2	237	0.2	1	1	0	0	1	3	3
107	4	8	1	225	0.1	0	1	0	0	1	2	3
108	5	10	1	157	0.1	0.5	1	0	0	1	2.5	3
109	6	11	1	188	0.1	1	1	0	0	1	3	3
110	5	7	1	220	0	0.5	0.5	0	0	1	2	3
111	3	5	2	231	0.1	0	0	0	0	1	1	
112	8	12	1	169	0.1	1	1	0	0	1	3	3
113	20	32	2	219	0.3	1	1	0	0	0	2	3
114	6	10	1	172	0.1	1	1	0	0	1	3	3
115	9	15	2	196	0.2	1	1	0	0	1	3	16
116	3	5	5	337	0.6	0	0	1	0	0	1	16
117	11	20	3	118	0.5	1	1	1	1	0	4	
118	5	9	11	336	0.1	0.5	1	0	0	1	2.5	10
119	19	33	5	285	0.2	1	1	1	0	1	4	5
120	3	6	6	325	0.1	0	0	0	0	1	1	
121	7	12	17	335	0.2	1	1	0	0	1	3	
122	4	8	3	268	0.1	0	1	1	0	1	3	
123	5	7	2	243	0.1	0.5	0.5	0	0	1	2	3
124	3	4	9	339	0.1	0	0	0	0	1	1	
125	6	10	12	326	0.2	1	1	0	0	1	3	10
126	13	18	13	314	0.2	1	1	0	0	1	3	17
127	4	7	6	312	0.2	0	0.5	0	0	1	1.5	
128	12	19	10	258	0.6	1	1	1	0	0	3	
129	8	12	11	331	0.4	1	1	0	0	0	2	

Appendix G

Cross-correlation

Input file used for the cross-correlation (program CORR) in SeisAn (Figures [G.1](#)[G.2](#)). Note: SFILE INDEXFILE correponds to a file containing a list of events. Here the list consists in all events identified as Local quakes (LQ) or Local uncertain (LU). Figures [G.3](#)[G.6](#) shows the output of CORR and Figures [G.7](#)[G.9](#) shows the output file after running XCLUST in SeisAN, showing the clusters and their events. Events are numbered and correspond to the ID of the event in Appendix [E](#). Clusters are also shown on Figure [G.10](#)

```

#
# corr input file

KEYWORD.....Comments.....Par 1.....Par 2.....

SFILE INDEXFILE                all_LQLU.lis

# minimum correlation value
MIN CORR                        0.7

# number of stations required for event pair to be correlated
MIN CORR CHAN                   1

# either 0. or 1.
INTERACTIVE                     0.

# activate (1.) or deactivate (0.) filters set below
FILTER                          1.

# time of window before phase as ratio of selected duration
PRE SIGNAL                      0.1

# use either full trace (0.) or same criteria as for master event (1.)
EVENT SELCRIT                   1.

# double sample rate n times
N DOUBLE SRATE                  0.

# time window for cross-correlation matrix, 999. for max
#CC MATRIX WINDOW              4.

# set hypocenter start latitude and longitude, 999. to not use
# (only affects output file, and use of location program afterwards)
START LATITUDE                  999.
START LONGITUDE                 999.

# set hypocenter fixed depth, 999. to not use
# (only affects output file, and use of location program afterwards)
FIX DEPTH                      999.

# keep original waveform names (0.) or put corr output file names (1.)
WAVENAME OUT                    0.

# output of correlation function, 0 for none,
# 1. for full data, 2 for reduced
# data 1 for data > MIN CORRELATION, otherwise 0
WAVE CORR OUT                   0.

# reduce to 0 (-) and 1 (+) if set to 1, full data if set to 1.
SINGLE BIT                       0.

# write out phases for correlation above threshold if set to 1.
CONTINUOUS MODE                 0.

```

Figure G.1: Input file for cross-correlation using CORR in SeisAn. Page 1/2


```

# extract event files for correlation above threshold, 0. for no extract,
# 1. to extract single channel used, 2. extract all channels
CONTINUOUS EXTRACT          0.

# flag to write corr.trace output file
TRACE OUTPUT                1.

# maximum distance between event pair to compute correlation
MAX EVENT DISTANCE          5.

# maximum distance between event and station
MAX STAT DISTANCE           50.

# setup of stations
#  STAT, COMP    - station and component codes
#  SELCRIT       - 1=P, 2=S, 4=full trace
#  DURATION      - signal duration in seconds if selcrit<4
#  FLOW, FHIGH   - filter limits for bandpass filter, can be
#                  disabled by FILTER (see above)
-----
KEYWORD...STAT.....COMP.....SELCRIT...DURATION..FLOW.....FHIGH.....

--- p ---
STATION  FER1      D Z      1.      4.      1.      30.
STATION  FER2      D Z      1.      4.      1.      30.
STATION  FER3      D Z      1.      4.      1.      30.
STATION  FER4      D Z      1.      4.      1.      30.
STATION  FER5      D Z      1.      4.      1.      30.
STATION  FER7      D Z      1.      4.      1.      30.
STATION  LAN1      D Z      1.      4.      1.      30.
STATION  LAN2      D Z      1.      4.      1.      30.
STATION  LAN3      D Z      1.      4.      1.      30.
STATION  LAN5      D Z      1.      4.      1.      30.
STATION  LAN6      D Z      1.      4.      1.      30.
STATION  LAN8      D Z      1.      4.      1.      30.
STATION  LAN9      D Z      1.      4.      1.      30.
STATION  LAN10     D Z      1.      4.      1.      30.
STATION  BOU4      D Z      1.      4.      1.      30.
STATION  BOU2      D Z      1.      4.      1.      30.
STATION  BOU3      D Z      1.      4.      1.      30.
STATION  EST1      D Z      1.      4.      1.      30.
STATION  EST2      D Z      1.      4.      1.      30.
STATION  EST3      D Z      1.      4.      1.      30.
--- s ---
STATION  FER1      D E      2.      4.      1.      30.
STATION  FER2      D E      2.      4.      1.      30.
STATION  FER3      D E      2.      4.      1.      30.
STATION  FER4      D E      2.      4.      1.      30.
STATION  FER5      D E      2.      4.      1.      30.
STATION  FER7      D E      2.      4.      1.      30.
STATION  LAN1      D E      2.      4.      1.      30.
STATION  LAN2      D E      2.      4.      1.      30.
STATION  LAN3      D E      2.      4.      1.      30.
STATION  LAN5      D E      2.      4.      1.      30.

```

Figure G.2: Input file for cross-correlation using CORR in SeisAn. Page 2/2

2 02-2045-37L.S201908	5 06-0554-53L.S201910	13 0.829
2 02-2045-37L.S201908	7 08-1834-33L.S201910	9 0.830
2 02-2045-37L.S201908	8 14-0436-14L.S201910	3 0.839
3 12-2136-22L.S201909	113 27-1055-08L.S202106	1 0.711
5 06-0554-53L.S201910	7 08-1834-33L.S201910	10 0.885
5 06-0554-53L.S201910	8 14-0436-14L.S201910	3 0.839
6 06-0720-52L.S201910	12 25-1123-55L.S201911	5 0.837
7 08-1834-33L.S201910	8 14-0436-14L.S201910	3 0.810
12 25-1123-55L.S201911	26 05-1514-03L.S202006	1 0.710
14 27-0218-48L.S202001	15 05-1146-35L.S202002	12 0.849
14 27-0218-48L.S202001	18 14-0255-33L.S202002	18 0.937
15 05-1146-35L.S202002	18 14-0255-33L.S202002	17 0.899
19 15-0500-44L.S202002	20 15-0501-49L.S202002	9 0.976
19 15-0500-44L.S202002	59 18-0309-18L.S202010	1 0.731
19 15-0500-44L.S202002	104 06-1321-57L.S202106	5 0.903
20 15-0501-49L.S202002	59 18-0309-18L.S202010	1 0.732
20 15-0501-49L.S202002	104 06-1321-57L.S202106	7 0.862
22 07-2007-29L.S202003	77 05-1412-09L.S202102	2 0.870
22 07-2007-29L.S202003	78 06-0244-09L.S202102	1 0.821
25 20-1727-47L.S202005	33 12-0612-07L.S202007	16 0.897
25 20-1727-47L.S202005	41 26-0052-47L.S202007	4 0.820
28 04-1758-49L.S202007	29 06-1200-36L.S202007	8 0.848
28 04-1758-49L.S202007	31 08-0633-23L.S202007	10 0.808
29 06-1200-36L.S202007	31 08-0633-23L.S202007	8 0.863
30 07-1103-35L.S202007	48 28-0619-18L.S202009	7 0.818
30 07-1103-35L.S202007	49 28-0622-57L.S202009	7 0.823
30 07-1103-35L.S202007	50 28-0627-26L.S202009	9 0.834
30 07-1103-35L.S202007	53 15-0011-50L.S202010	5 0.887
30 07-1103-35L.S202007	54 15-0016-55L.S202010	8 0.857
30 07-1103-35L.S202007	55 15-0018-36L.S202010	4 0.866
30 07-1103-35L.S202007	56 15-0022-00L.S202010	7 0.860
30 07-1103-35L.S202007	57 15-0059-02L.S202010	7 0.860
30 07-1103-35L.S202007	58 15-0101-56L.S202010	5 0.843
30 07-1103-35L.S202007	61 20-1752-31L.S202010	5 0.789
30 07-1103-35L.S202007	62 20-1806-18L.S202010	5 0.794
30 07-1103-35L.S202007	63 20-1836-41L.S202010	3 0.760
30 07-1103-35L.S202007	64 20-1843-10L.S202010	2 0.745
32 11-0526-00L.S202007	34 12-1231-25L.S202007	18 0.844
32 11-0526-00L.S202007	35 12-1313-28L.S202007	7 0.776
32 11-0526-00L.S202007	36 12-1343-35L.S202007	9 0.825
32 11-0526-00L.S202007	38 15-0301-04L.S202007	15 0.850
33 12-0612-07L.S202007	41 26-0052-47L.S202007	7 0.831
34 12-1231-25L.S202007	35 12-1313-28L.S202007	8 0.807
34 12-1231-25L.S202007	36 12-1343-35L.S202007	11 0.798
34 12-1231-25L.S202007	38 15-0301-04L.S202007	33 0.844
34 12-1231-25L.S202007	119 30-2350-31L.S202106	1 0.727
35 12-1313-28L.S202007	36 12-1343-35L.S202007	12 0.911
35 12-1313-28L.S202007	38 15-0301-04L.S202007	11 0.897
36 12-1343-35L.S202007	38 15-0301-04L.S202007	12 0.938
36 12-1343-35L.S202007	119 30-2350-31L.S202106	1 0.722
48 28-0619-18L.S202009	49 28-0622-57L.S202009	13 0.939
48 28-0619-18L.S202009	50 28-0627-26L.S202009	12 0.940
48 28-0619-18L.S202009	53 15-0011-50L.S202010	3 0.924
48 28-0619-18L.S202009	54 15-0016-55L.S202010	6 0.870

Figure G.3: Output file after running CORR in SeisAn. Page 1/4

48 28-0619-18L.S202009	55 15-0018-36L.S202010	7 0.823
48 28-0619-18L.S202009	56 15-0022-00L.S202010	7 0.834
48 28-0619-18L.S202009	57 15-0059-02L.S202010	8 0.843
48 28-0619-18L.S202009	58 15-0101-56L.S202010	3 0.854
48 28-0619-18L.S202009	61 20-1752-31L.S202010	5 0.786
48 28-0619-18L.S202009	62 20-1806-18L.S202010	5 0.777
48 28-0619-18L.S202009	63 20-1836-41L.S202010	3 0.733
49 28-0622-57L.S202009	50 28-0627-26L.S202009	12 0.963
49 28-0622-57L.S202009	53 15-0011-50L.S202010	3 0.921
49 28-0622-57L.S202009	54 15-0016-55L.S202010	6 0.871
49 28-0622-57L.S202009	55 15-0018-36L.S202010	8 0.829
49 28-0622-57L.S202009	56 15-0022-00L.S202010	6 0.845
49 28-0622-57L.S202009	57 15-0059-02L.S202010	9 0.851
49 28-0622-57L.S202009	58 15-0101-56L.S202010	3 0.852
49 28-0622-57L.S202009	61 20-1752-31L.S202010	5 0.827
49 28-0622-57L.S202009	62 20-1806-18L.S202010	5 0.816
49 28-0622-57L.S202009	63 20-1836-41L.S202010	3 0.781
50 28-0627-26L.S202009	53 15-0011-50L.S202010	7 0.895
50 28-0627-26L.S202009	54 15-0016-55L.S202010	11 0.892
50 28-0627-26L.S202009	55 15-0018-36L.S202010	10 0.844
50 28-0627-26L.S202009	56 15-0022-00L.S202010	13 0.846
50 28-0627-26L.S202009	57 15-0059-02L.S202010	14 0.853
50 28-0627-26L.S202009	58 15-0101-56L.S202010	7 0.842
50 28-0627-26L.S202009	61 20-1752-31L.S202010	10 0.830
50 28-0627-26L.S202009	62 20-1806-18L.S202010	12 0.833
50 28-0627-26L.S202009	63 20-1836-41L.S202010	7 0.807
50 28-0627-26L.S202009	64 20-1843-10L.S202010	1 0.772
53 15-0011-50L.S202010	54 15-0016-55L.S202010	12 0.932
53 15-0011-50L.S202010	55 15-0018-36L.S202010	8 0.903
53 15-0011-50L.S202010	56 15-0022-00L.S202010	13 0.941
53 15-0011-50L.S202010	57 15-0059-02L.S202010	12 0.915
53 15-0011-50L.S202010	58 15-0101-56L.S202010	7 0.863
53 15-0011-50L.S202010	61 20-1752-31L.S202010	3 0.846
53 15-0011-50L.S202010	62 20-1806-18L.S202010	5 0.862
53 15-0011-50L.S202010	63 20-1836-41L.S202010	2 0.824
53 15-0011-50L.S202010	64 20-1843-10L.S202010	2 0.809
54 15-0016-55L.S202010	55 15-0018-36L.S202010	7 0.910
54 15-0016-55L.S202010	56 15-0022-00L.S202010	10 0.942
54 15-0016-55L.S202010	57 15-0059-02L.S202010	16 0.879
54 15-0016-55L.S202010	58 15-0101-56L.S202010	9 0.847
54 15-0016-55L.S202010	61 20-1752-31L.S202010	4 0.814
54 15-0016-55L.S202010	62 20-1806-18L.S202010	7 0.840
54 15-0016-55L.S202010	63 20-1836-41L.S202010	2 0.830
54 15-0016-55L.S202010	64 20-1843-10L.S202010	2 0.820
55 15-0018-36L.S202010	56 15-0022-00L.S202010	9 0.898
55 15-0018-36L.S202010	57 15-0059-02L.S202010	12 0.929
55 15-0018-36L.S202010	58 15-0101-56L.S202010	9 0.881
55 15-0018-36L.S202010	61 20-1752-31L.S202010	7 0.828
55 15-0018-36L.S202010	62 20-1806-18L.S202010	9 0.822
55 15-0018-36L.S202010	63 20-1836-41L.S202010	2 0.810
56 15-0022-00L.S202010	57 15-0059-02L.S202010	12 0.898
56 15-0022-00L.S202010	58 15-0101-56L.S202010	8 0.840
56 15-0022-00L.S202010	61 20-1752-31L.S202010	4 0.805
56 15-0022-00L.S202010	62 20-1806-18L.S202010	7 0.823

Figure G.4: Output file after running CORR in SeisAn. Page 2/4

56 15-0022-00L.S202010	63 20-1836-41L.S202010	3 0.765
56 15-0022-00L.S202010	64 20-1843-10L.S202010	2 0.776
57 15-0059-02L.S202010	58 15-0101-56L.S202010	11 0.921
57 15-0059-02L.S202010	61 20-1752-31L.S202010	10 0.883
57 15-0059-02L.S202010	62 20-1806-18L.S202010	14 0.842
57 15-0059-02L.S202010	63 20-1836-41L.S202010	5 0.846
57 15-0059-02L.S202010	64 20-1843-10L.S202010	2 0.867
58 15-0101-56L.S202010	61 20-1752-31L.S202010	7 0.915
58 15-0101-56L.S202010	62 20-1806-18L.S202010	9 0.866
58 15-0101-56L.S202010	63 20-1836-41L.S202010	6 0.803
58 15-0101-56L.S202010	64 20-1843-10L.S202010	3 0.830
61 20-1752-31L.S202010	62 20-1806-18L.S202010	20 0.957
61 20-1752-31L.S202010	63 20-1836-41L.S202010	16 0.897
61 20-1752-31L.S202010	64 20-1843-10L.S202010	7 0.799
62 20-1806-18L.S202010	63 20-1836-41L.S202010	17 0.886
62 20-1806-18L.S202010	64 20-1843-10L.S202010	4 0.852
63 20-1836-41L.S202010	64 20-1843-10L.S202010	6 0.952
71 14-2209-24L.S202012	72 15-0141-04L.S202012	10 0.914
71 14-2209-24L.S202012	87 24-1807-16L.S202102	13 0.857
72 15-0141-04L.S202012	87 24-1807-16L.S202102	6 0.799
77 05-1412-09L.S202102	78 06-0244-09L.S202102	2 0.764
79 07-0937-19L.S202102	126 21-0301-40L.S202107	1 0.731
80 10-1605-09L.S202102	81 11-0539-51L.S202102	9 0.904
80 10-1605-09L.S202102	84 13-1114-21L.S202102	2 0.780
81 11-0539-51L.S202102	84 13-1114-21L.S202102	3 0.794
82 11-2221-59L.S202102	83 11-2245-41L.S202102	7 0.907
88 07-1531-57L.S202103	89 07-1541-16L.S202103	14 0.978
88 07-1531-57L.S202103	91 14-0245-37L.S202103	16 0.908
88 07-1531-57L.S202103	92 14-0517-32L.S202103	12 0.914
88 07-1531-57L.S202103	93 14-0532-46L.S202103	13 0.808
88 07-1531-57L.S202103	95 18-0507-38L.S202103	13 0.793
88 07-1531-57L.S202103	96 25-1249-40L.S202103	11 0.860
89 07-1541-16L.S202103	91 14-0245-37L.S202103	15 0.916
89 07-1541-16L.S202103	92 14-0517-32L.S202103	15 0.919
89 07-1541-16L.S202103	93 14-0532-46L.S202103	13 0.804
89 07-1541-16L.S202103	95 18-0507-38L.S202103	16 0.800
89 07-1541-16L.S202103	96 25-1249-40L.S202103	13 0.850
89 07-1541-16L.S202103	103 05-1725-36L.S202106	3 0.741
91 14-0245-37L.S202103	92 14-0517-32L.S202103	13 0.942
91 14-0245-37L.S202103	93 14-0532-46L.S202103	15 0.872
91 14-0245-37L.S202103	95 18-0507-38L.S202103	19 0.902
91 14-0245-37L.S202103	96 25-1249-40L.S202103	15 0.904
91 14-0245-37L.S202103	103 05-1725-36L.S202106	1 0.745
92 14-0517-32L.S202103	93 14-0532-46L.S202103	12 0.908
92 14-0517-32L.S202103	95 18-0507-38L.S202103	15 0.841
92 14-0517-32L.S202103	96 25-1249-40L.S202103	12 0.874
92 14-0517-32L.S202103	103 05-1725-36L.S202106	3 0.751
93 14-0532-46L.S202103	95 18-0507-38L.S202103	15 0.881
93 14-0532-46L.S202103	96 25-1249-40L.S202103	7 0.857
95 18-0507-38L.S202103	96 25-1249-40L.S202103	15 0.814
95 18-0507-38L.S202103	103 05-1725-36L.S202106	1 0.785
96 25-1249-40L.S202103	103 05-1725-36L.S202106	1 0.798
101 21-0152-55L.S202105	102 22-1505-27L.S202105	14 0.781
101 21-0152-55L.S202105	115 28-0452-09L.S202106	7 0.738

Figure G.5: Output file after running CORR in SeisAn. Page 3/4

102 22-1505-27L.S202105	115 28-0452-09L.S202106	7 0.788
102 22-1505-27L.S202105	116 28-1318-32L.S202106	2 0.785
105 07-1611-22L.S202106	125 17-2245-25L.S202107	7 0.898
106 09-0830-06L.S202106	107 11-1308-03L.S202106	3 0.812
106 09-0830-06L.S202106	108 12-2240-03L.S202106	4 0.819
106 09-0830-06L.S202106	109 15-1418-05L.S202106	1 0.908
106 09-0830-06L.S202106	110 15-1521-55L.S202106	3 0.796
106 09-0830-06L.S202106	112 15-2304-01L.S202106	5 0.800
106 09-0830-06L.S202106	113 27-1055-08L.S202106	6 0.783
106 09-0830-06L.S202106	114 27-1712-47L.S202106	1 0.763
106 09-0830-06L.S202106	123 13-1115-56L.S202107	1 0.925
107 11-1308-03L.S202106	108 12-2240-03L.S202106	2 0.859
107 11-1308-03L.S202106	109 15-1418-05L.S202106	3 0.795
107 11-1308-03L.S202106	110 15-1521-55L.S202106	3 0.820
107 11-1308-03L.S202106	112 15-2304-01L.S202106	8 0.797
107 11-1308-03L.S202106	113 27-1055-08L.S202106	1 0.881
107 11-1308-03L.S202106	114 27-1712-47L.S202106	1 0.734
107 11-1308-03L.S202106	123 13-1115-56L.S202107	1 0.878
108 12-2240-03L.S202106	109 15-1418-05L.S202106	2 0.815
108 12-2240-03L.S202106	110 15-1521-55L.S202106	2 0.794
108 12-2240-03L.S202106	112 15-2304-01L.S202106	1 0.902
108 12-2240-03L.S202106	113 27-1055-08L.S202106	3 0.762
108 12-2240-03L.S202106	114 27-1712-47L.S202106	2 0.730
108 12-2240-03L.S202106	123 13-1115-56L.S202107	1 0.850
109 15-1418-05L.S202106	110 15-1521-55L.S202106	6 0.932
109 15-1418-05L.S202106	112 15-2304-01L.S202106	11 0.885
109 15-1418-05L.S202106	113 27-1055-08L.S202106	1 0.919
109 15-1418-05L.S202106	114 27-1712-47L.S202106	1 0.747
109 15-1418-05L.S202106	123 13-1115-56L.S202107	2 0.844
110 15-1521-55L.S202106	112 15-2304-01L.S202106	6 0.964
110 15-1521-55L.S202106	113 27-1055-08L.S202106	1 0.907
110 15-1521-55L.S202106	123 13-1115-56L.S202107	2 0.805
112 15-2304-01L.S202106	113 27-1055-08L.S202106	3 0.876
112 15-2304-01L.S202106	123 13-1115-56L.S202107	1 0.899
113 27-1055-08L.S202106	114 27-1712-47L.S202106	1 0.762
113 27-1055-08L.S202106	123 13-1115-56L.S202107	2 0.872
118 30-2349-02L.S202106	125 17-2245-25L.S202107	6 0.776

Figure G.6: Output file after running CORR in SeisAn. Page 4/4

```

group:      1 number of events:      8
-----
event links
-----
    88      6
    89      7
    92      7
    91      7
    93      6
    96      7
    95      7
   103      5
=====
group:      2 number of events:      4
-----
event links
-----
    19      3
    20      3
   104      2
    59      2
=====
group:      3 number of events:     10
-----
event links
-----
   110      7
   112      7
   109      8
   113      9
   106      8
   108      8
   123      7
   107      8
   114      5
     3      1
=====
group:      4 number of events:     14
-----
event links
-----
    49     12
    50     13
    48     12
    53     13
    57     13
    55     12
    56     13
    54     13
    30     13
    61     13
    58     13
    64     10
    62     13
    63     13
=====
group:      5 number of events:      6
-----
event links
-----
    36      5
    38      4
    35      4
    32      4
    34      5
   119      2
=====

```

Figure G.7: Output file after running XCLUST in SeisAn. Page 1/3

```

group:      6 number of events:      3
-----
event links
-----
    14      2
    18      2
    15      2
=====
group:      7 number of events:      3
-----
event links
-----
    71      2
    72      2
    87      2
=====
group:      8 number of events:      2
-----
event links
-----
    82      1
    83      1
=====
group:      9 number of events:      3
-----
event links
-----
    80      2
    81      2
    84      2
=====
group:     10 number of events:      3
-----
event links
-----
   105      1
   125      2
   118      1
=====
group:     11 number of events:      3
-----
event links
-----
    25      2
    33      2
    41      2
=====
group:     12 number of events:      4
-----
event links
-----
     5      3
     7      3
     8      3
     2      3
=====
group:     13 number of events:      3
-----
event links
-----
    22      2
    77      2
    78      2
=====
group:     14 number of events:      3
-----
event links

```

Figure G.8: Output file after running XCLUST in SeisAn. Page 2/3

```

-----
    29    2
    31    2
    28    2
=====
group:   15 number of events:    3
-----
event links
-----
    6    1
   12    2
   26    1
=====
group:   16 number of events:    4
-----
event links
-----
   102    3
   115    2
   116    1
   101    2
=====
group:   17 number of events:    2
-----
event links
-----
    79    1
   126    1

```

Figure G.9: Output file after running XCLUST in SeisAn. Page 3/3

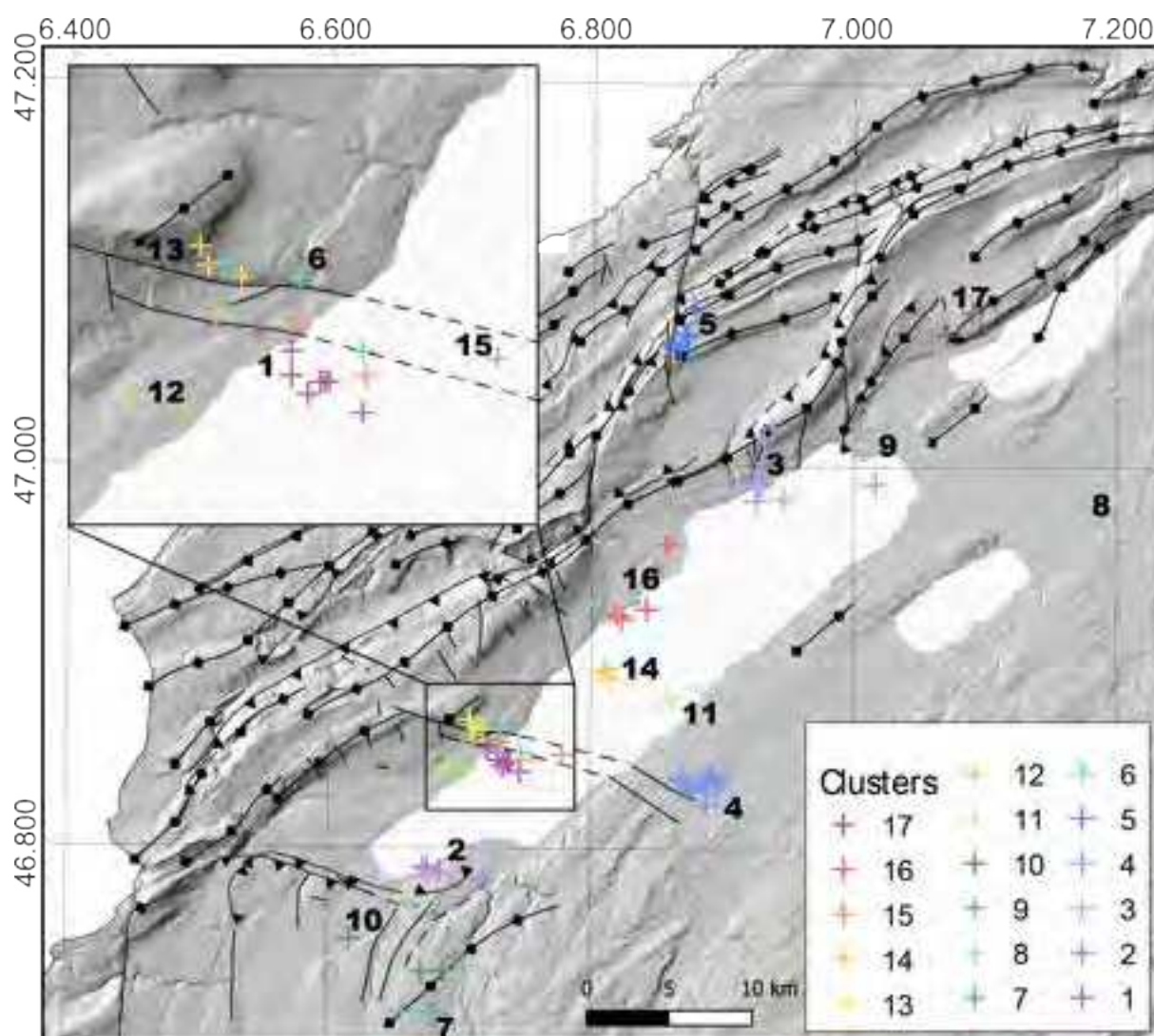
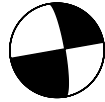


Figure G.10: Clusters identified after running the program XCLUST in SeisAn.

Appendix H

HASH : input and output

Focal mechanism cluster 1



Data input

Number of polarities = 28

Allowed polarity errors = 2

Degree increment of P-T axes = 5

Data output

Strike,dip,rake 80.0 90.0 -165.0

Fault+aux plane uncertainty 20.1 13.9

Weighted fraction of pol misfits 0.08

Average amplitude error 0.00

Station dist ratio 0.75

Focal mechanism cluster 3



Data input

Number of polarities = 36

Allowed polarity errors = 4

Degree increment of P-T axes for search = 5

Data output

Strike,dip,rake 234.0 25.0 -160.0

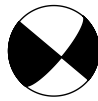
Fault+aux plane uncertainty 23.0 26.0

Weighted fraction of pol misfits 0.14

Average amplitude error 0.00

Station dist ratio 0.65

Focal mechanism cluster 4



Data input

Number of polarities = 31

Allowed polarity errors = 1

Degree increment of P-T axes for search = 10

Data output

Strike,dip,rake 310.0 90.0 160.0

Fault+aux plane uncertainty 27.3 21.0

Weighted fraction of pol misfits 0.01

Average amplitude error 0.00

Station dist ratio 0.63

Focal mechanism cluster 5



Data input

Number of polarities = 40

Allowed polarity errors = 5

Degree increment of P-T axes for search = 5

Data output

Strike,dip,rake 253.6 50.0 150.0

Fault+aux plane uncertainty 39.4 20.6

Weighted fraction of pol misfits 0.15

Average amplitude error 0.00

Station dist ratio 0.62

Focal mechanism cluster 6



Data input

Number of polarities = 16

Allowed polarity errors = 1

Degree increment of P-T axes for search = 5

Data output

Strike,dip,rake 25.0 90.0 -160.0

Fault+aux plane uncertainty 7.5 19.1

Weighted fraction of pol misfits 0.07

Average amplitude error 0.00

Station dist ratio 0.65

Figure H.1: Input and output data for and from running HASH.

Appendix I

Location uncertainty

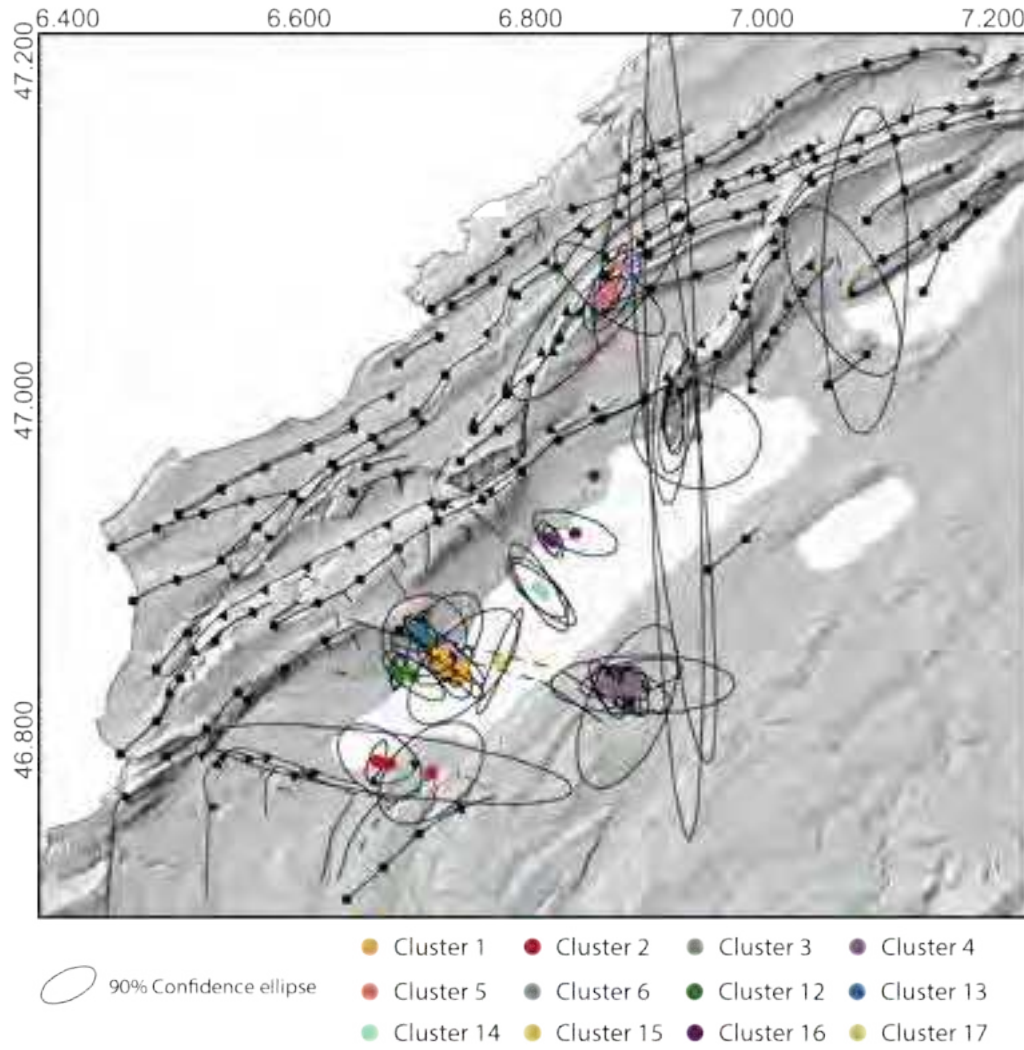


Figure I.1: 90% confidence ellipse for earthquakes of clusters 1-6 and 12-17.

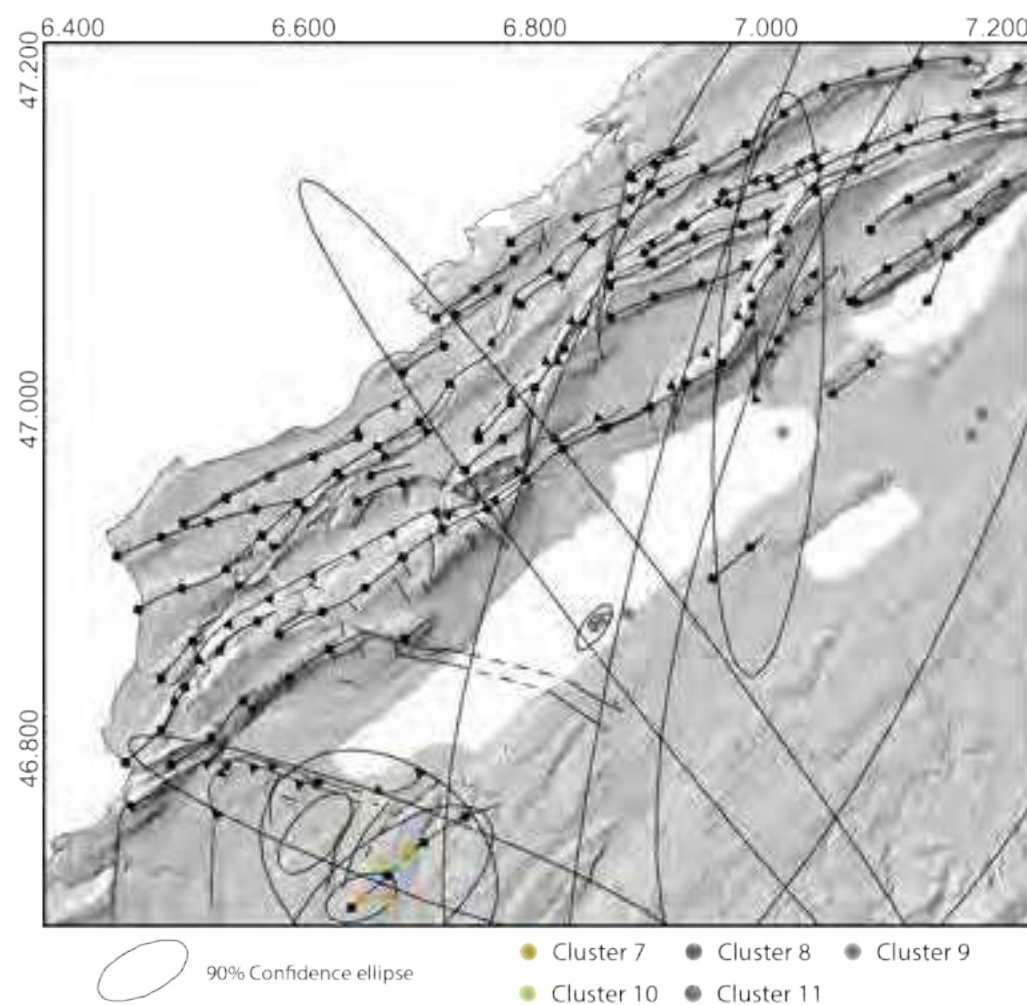


Figure I.2: 90% confidence ellipse for earthquakes of clusters 7-11.

Appendix J

HypoDD

Input and output files for and from hypoDD (Figures [J.1](#)[J.4](#)) and relocated events (Figure [J.5](#)).

```

hypoDD_2
* RELOC.INP:
*--- input file selection
dt.cc
*
*catalog P diff times:
dt.ct
*
* event file:
event.dat
*
* station file:
station_alt.dat
*
*--- output file selection
* original locations:
hypoDD.loc
* relocations:
hypoDD.reloc
* station information:
hypoDD.sta
* residual information:
hypoDD.res
* source parameter information:
hypoDD.src
*
*
*--- data type selection:
* IDAT: 1= cross corr; 2= catalog; 3= cross & cat
* IPHA: 1= P; 2= S; 3= P&S
* DIST:max dist [km] between cluster centroid and station
* IDAT  IPHA  MAXDIST
   3      3      120
*
*--- event clustering:
* OBSCC:  min # of obs/pair for crosstime data (0= no clustering)
* OBSCT:  min # of obs/pair for network data (0= no clustering)
* MINDIST: min pair-station distance (-9= not active)
* MAXDIST: max pair-station distance (-9= not active)
* MAXGAP:  max azimuthal gap (-9= not active)
* OBSCC  OBSCT  MINDIST  MAXDIST  MAXGAP
   3      10      -999      -999      -999
*
*--- solution control:
* ISTART:      1 = from single source; 2 = from network sources
* ISOLV:       1 = SVD, 2=lsqr
* IAQ:         remove airquakes: 0=NO, 1=YES
* NSET:        number of sets of iteration with specifications following
* ISTART  ISOLV  IAQ  NSET
   2       1     0    4
*
*--- data weighting and re-weighting:
* NITER:       last iteration to used the following weights
* WTCCP, WTCCS: weight cross P, S

```

Figure J.1: Input file for hypoDD. Page 1/3

```

* WTCTP, WTCTS:      weight catalog P, S
* WRCC, WRCT:        residual threshold in sec for cross, catalog data
* WDCC, WDCT:        max dist [km] between cross, catalog linked pairs
* DAMP:              damping (for lsqr only)
*
* --- CROSS DATA ----- CATALOG DATA ----
* NITER  WTCCP  WTCCS  WRCC  WDCC  WTCTP  WTCTS  WRCT  WDCT  DAMP
* 1      0.01  0.005  -999  -999  1.00   0.5   -9   5   20  ctcc
* 3      0.01  0.005  -999  -999  1.00   0.5   10   5   20  ctcc
* 3      1     0.5    10    5    1.00   0.5    9   5   20  ctcc
* 3      1     0.5    9     5    0.1    0.05   9   5   20  ctcc
*
* 1      1     0.5   -999  -999  1.00   0.5  -999  -999  20  cc
* 3      1     0.5    5     3    1.00   0.5   10   10   20  cc
* 3      1     0.5    4     2    1.00   0.5   10   10   20  cc
*
* 1      1     0.5   -999  -999  1.00   0.5   40  -999  20  ct
* 5      1     0.5   10     5    1.00   0.5   20   5   20  ct
* 5      1     0.5   10     5    1.00   0.5   11   3   20  ct
*
* --- FORWARD MODEL:
* IMOD  0= 1D local, layered, P velocity w/ fixed vp/vs ratio (hypoDD1.0);
*        1= 1D local, layered, P velocity w/ variable vp/vs ratio
*        9= 3D local model (simulps raytracer)
0
* --- 1D model, fixed vp/vs ratio:
* NLAY:      number of model layers
* RATIO:      vp/vs ratio
* TOP:        depths of top of layer (km)
* VEL:        layer velocities (km/s)
* IMOD
* NLAY  RATIO
* MOD_TOP , MOD_V
11      1.71
-5.00 0.00 2.00 5.00 8.00 15.00 22.00 30.00 40.00 50.0 60.0
4.60 4.60 5.98 6.00 6.03 6.09 6.16 6.24 6.33 8.10 8.15
***0.00 0.25 1.50 2.50 3.50 5.00 6.00 9.00 15.00 25.00
**1.42 3.24 4.82 5.36 5.60 5.65 5.90 6.15 6.60 8.00
*
*
* --- 1D model, variable vp/vs ratio:
* TOP:        depths of top of layer (km)
* VEL:        layer velocities (km/s) end w/ -9
* RATIO:      layer ratios end w/ -9
* IMOD
* 1
* 0.00 0.25 1.50 2.50 3.50 5.00 6.00 9.00 15.00 25.00 -9
* 1.42 3.24 4.82 5.36 5.60 5.65 5.90 6.15 6.60 8.00 -9
* 1.73 1.73 1.73 1.73 1.73 1.73 1.73 1.73 1.73 1.73 -9
*
* --- Local 3D model:
* IMOD
* 3DMOD      name of 3D velocity model
* LAT_3D LON_3D ROT_3D origin of 3D velocity model and rotation of
*                  cart coord. system (pos= anti clockwise)

```

Figure J.2: Input file for hypoDD. Page 2/3

```

*                               raytracing parameters
* IPHA NDIP ISKIP SCALE1 SCALE2 XFAX TLIM NITPB simulps ray tracing parameters
*9
*vel3D.dat
*37.878078 -122.243950 0
*2  9  4  1  1  1.35 0.0005 50
**2  9  2  1  1  1.35 0.0005 50
*
*--- event selection:
* CID: cluste 0r to be relocated (0 = all)
* ID:  cuspids of event to be relocated (8 per line)
* CID
  0
* ID

```

Figure J.3: Input file for hypoDD. Page 3/3

A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	P	Q	R	S	T	U	V	W	X
335	46.834835	6.883679	3.503	-30.8	36.6	1236.6	14.3	5.6	13.7	2020	9	28	6	19	18.580	0.30	14	51	39	40	0.002	0.048	1
37	46.835055	6.883143	3.526	-71.7	61.1	1259.7	9.5	5.2	12.1	2020	9	28	6	27	26.340	1.20	22	83	53	52	0.004	0.055	1
36	46.834847	6.883507	3.509	-44.0	38.0	1242.1	13.9	5.5	13.7	2020	9	28	6	22	57.140	0.90	14	51	36	40	0.002	0.028	1
40	46.834391	6.884570	3.473	37.2	-12.7	1206.0	9.6	7.0	15.9	2020	10	15	0	11	49.990	0.50	4	63	30	46	0.002	0.048	1
41	46.834469	6.884182	3.488	7.5	-4.1	1221.5	8.4	5.9	11.4	2020	10	15	0	16	55.600	1.20	18	65	33	43	0.003	0.048	1
43	46.834493	6.884290	3.492	15.8	-1.4	1224.9	9.9	6.3	12.1	2020	10	15	0	22	0.190	0.60	22	61	32	42	0.005	0.050	1
44	46.834147	6.885172	3.459	83.0	-39.8	1191.8	7.9	5.0	10.3	2020	10	15	0	59	2.300	0.90	33	78	45	51	0.004	0.047	1
45	46.833667	6.885329	3.432	95.1	-93.2	1165.6	12.3	7.6	14.7	2020	10	15	1	56.710	0.60	13	57	24	41	0.003	0.046	1	
47	46.833464	6.885572	3.388	113.6	-115.8	1120.9	11.0	7.6	13.8	2020	10	20	17	52	31.460	1.10	25	59	50	43	0.003	0.050	1
48	46.833321	6.885825	3.375	132.9	-131.6	1108.2	10.3	7.5	13.5	2020	10	20	18	6	18.520	1.80	30	71	57	48	0.005	0.066	1
49	46.832959	6.886314	3.330	170.2	-171.9	1063.4	12.6	8.8	20.2	2020	10	20	18	36	41.620	0.70	17	47	41	42	0.004	0.051	1
69	46.845736	6.732973	2.655	91.2	-177.8	43.6	14.0	14.1	29.0	2021	3	7	15	31	57.900	0.10	21	58	31	38	0.002	0.047	2
70	46.845695	6.733039	2.657	96.2	-182.3	45.6	12.5	10.8	27.6	2021	3	7	15	41	15.990	0.20	27	61	41	39	0.002	0.064	2
72	46.845825	6.733296	2.657	115.8	-167.8	45.8	12.9	12.7	26.7	2021	3	14	2	45	37.750	1.10	33	60	49	39	0.002	0.061	2
73	46.845988	6.733141	2.654	104.0	-149.7	43.3	13.4	11.1	28.0	2021	3	14	5	17	32.930	1.50	27	53	39	34	0.002	0.038	2
74	46.846456	6.733300	2.641	116.1	-97.7	30.0	14.5	14.5	28.0	2021	3	14	5	32	46.040	0.60	21	54	35	34	0.002	0.058	2
75	46.846322	6.733359	2.670	120.7	-112.6	59.2	12.2	11.0	26.6	2021	3	18	5	7	38.150	0.50	31	63	55	44	0.002	0.052	2
76	46.845459	6.733094	2.638	100.4	-208.5	26.9	12.5	11.0	27.7	2021	3	25	12	49	40.560	1.10	26	48	48	33	0.003	0.051	2
79	46.858647	6.721606	2.059	-775.7	1257.5	-551.6	62.5	67.8	123.9	2021	5	2	9	45	56.620	1.20	0	0	59	36	-9.000	0.062	2
82	46.845858	6.732227	2.869	34.3	-164.2	258.3	46.0	28.6	21.3	2021	6	5	17	25	36.360	1.80	0	5	63	41	0.006	0.064	2
23	47.068575	6.869453	3.417	-26.3	-2.3	-3.0	10.0	10.4	12.2	2020	7	11	5	26	0.280	0.30	21	28	21	15	0.002	0.067	4
25	47.068563	6.869807	3.447	0.6	-3.6	26.9	8.6	8.6	16.2	2020	7	12	12	31	25.390	2.50	26	44	29	32	0.002	0.055	4
26	47.068778	6.870063	3.398	20.0	20.4	-21.9	8.5	11.6	17.8	2020	7	12	13	13	28.120	0.80	14	24	4	6	0.002	0.088	4
27	47.068526	6.869817	3.423	1.3	-7.7	3.2	7.7	10.9	17.5	2020	7	12	13	43	35.780	0.40	14	30	13	24	0.002	0.075	4
29	47.068563	6.869856	3.415	4.2	-3.6	-5.2	6.6	8.3	11.5	2020	7	15	3	1	4.540	1.00	29	42	29	29	0.002	0.050	4
5	46.842643	6.761052	9.811	1376.8	-1284.7	892.1	152.2	230.4	309.3	2019	10	6	7	20	52.480	0.60	1	4	14	16	0.001	0.077	5
9	46.844027	6.761317	9.850	1397.0	-1130.9	930.1	153.5	227.9	305.8	2019	11	25	11	23	54.920	0.30	1	4	13	20	0.001	0.047	5
11	46.861247	6.730843	8.315	-927.1	783.5	-604.6	99.0	146.2	203.5	2020	1	27	2	18	49.080	0.70	10	20	22	21	0.002	0.057	5
12	46.861523	6.730867	8.311	-925.3	814.2	-609.5	99.1	145.8	203.5	2020	2	5	11	46	35.120	0.90	11	18	25	22	0.002	0.045	5
14	46.861552	6.730936	8.311	-920.0	817.4	-608.6	98.8	145.1	203.4	2020	2	14	2	55	33.610	0.90	13	22	30	31	0.002	0.064	5
80	46.924886	6.824876	42.876	-237.9	284.1	42865.7	509.6	417.2	136.3	2021	5	21	1	52	55.370	2.50	9	12	0	0	0.014	-9.000	6
81	46.918823	6.830579	42.819	196.5	-389.9	42809.0	477.7	442.2	120.1	2021	5	22	15	5	27.580	0.80	8	13	0	0	0.009	-9.000	6
94	46.923291	6.828546	42.747	41.6	106.8	42736.9	739.0	535.5	144.8	2021	6	28	4	52	9.260	0.80	7	7	0	0	0.017	-9.000	6
19	46.878809	6.866380	7.211	3.6	-20.8	6604.0	14.8	16.3	22.6	2020	5	20	17	27	47.220	0.70	7	13	0	0	0.002	-9.000	7
24	46.878544	6.866549	7.163	16.5	-50.2	6556.8	13.8	12.9	18.8	2020	7	12	6	12	7.300	1.00	7	16	0	0	0.002	-9.000	7
32	46.879651	6.866070	7.194	-20.1	72.8	6587.8	23.0	23.0	29.2	2020	7	26	0	52	47.380	-0.40	2	9	0	0	0.001	-9.000	7
84	46.743555	6.644356	2.169	-11.0	6.3	68.9	70.8	78.1	445.7	2021	6	7	16	11	22.400	1.70	4	3	5	3	0.002	0.063	8
101	46.743445	6.644648	2.031	11.3	-5.9	-68.9	70.8	78.1	445.7	2021	7	17	22	45	25.600	0.10	4	3	5	3	0.002	0.063	8
15	46.789718	6.673991	1.734	-38.8	24.0	-15.6	53.8	85.0	101.7	2020	2	15	5	1	49.500	0.40	0	6	4	4	0.000	0.033	9
83	46.789286	6.675009	1.766	38.8	-24.0	15.6	53.8	85.0	101.7	2021	6	6	13	21	57.900	0.60	0	6	4	4	0.000	0.033	9
54	46.699491	6.651083	8.799	-31.9	54.7	49.5	42.2	53.0	125.5	2020	12	14	22	9	24.600	1.40	5	9	0	0	0.007	-9.000	10
68	46.698507	6.651915	8.701	31.7	-54.7	-49.5	42.2	53.0	125.5	2021	2	24	18	7	16.100	1.20	5	9	0	0	0.007	-9.000	10
2	46.842415	6.690923	1.290	32.2	-9.0	-10.3	9.9	12.1	16.4	2019	8	2	20	45	36.960	0.70	3	10	5	4	0.002	0.026	12
4	46.842582	6.690077	1.310	-32.3	9.5	10.3	9.9	12.1	16.4	2019	10	6	5	54	53.830	-0.40	3	10	5	4	0.002	0.026	12

Figure J.4: HypoDD output file. Page 1/1. A: ID, B: LAT, C: LONG, D: DEPTH, E: X, F: Y, G: Z, H: EX, I: EY, J: EZ, K: YEAR, L: MONTH, M: DAY, N: HOUR, O: MIN, P: SEC, Q: MAG, R: NCCP, S: NCCS, T: NCTP, U: NCTS, V: RCC, W: RCT, X: CID.

ID	Date and time (UTC)	Latitude	Longitude	Depth	ML	Cluster	Depth diff.
2	02.08.2019 20:45:37	46.842	6.691	1.3	0.7	12	0.1
5	06.10.2019 05:54:54	46.843	6.69	1.3	-0.4	12	-0.1
6	06.10.2019 07:20:53	46.843	6.761	9.8	0.6	15	1.8
12	25.11.2019 11:23:55	46.844	6.761	9.9	0.3	15	-1.9
14	27.01.2020 02:18:49	46.861	6.731	8.3	0.7	6	1.6
15	05.02.2020 11:46:35	46.862	6.731	8.3	0.9	6	-1.2
18	14.02.2020 02:55:34	46.862	6.731	8.3	0.9	6	-0.3
20	15.02.2020 05:01:50	46.79	6.674	1.7	0.4	2	-0.5
25	20.05.2020 17:27:47	46.879	6.866	7.2	0.7	11	7.2
32	11.07.2020 05:26:00	47.069	6.869	3.4	0.3	5	2.5
33	12.07.2020 06:12:07	46.879	6.867	7.2	1	11	7.2
34	12.07.2020 12:31:25	47.069	6.87	3.5	2.5	5	1.3
35	12.07.2020 13:13:28	47.069	6.87	3.4	0.8	5	-5.8
36	12.07.2020 13:43:36	47.069	6.87	3.4	0.4	5	-0.9
38	15.07.2020 03:01:05	47.069	6.87	3.4	1	5	2.9
41	26.07.2020 00:52:47	46.88	6.866	7.2	-0.4	11	5.4
48	28.09.2020 06:19:19	46.835	6.884	3.5	0.3	4	1.7
49	28.09.2020 06:22:57	46.835	6.884	3.5	0.9	4	0.8
50	28.09.2020 06:27:26	46.835	6.883	3.5	1.2	4	0.7
53	15.10.2020 00:11:50	46.834	6.885	3.5	0.5	4	2.2
54	15.10.2020 00:16:56	46.834	6.884	3.5	1.2	4	1.6
56	15.10.2020 00:22:00	46.834	6.884	3.5	0.6	4	0.8
57	15.10.2020 00:59:02	46.834	6.885	3.5	0.9	4	2.1
58	15.10.2020 01:01:57	46.834	6.885	3.4	0.6	4	-0.6
61	20.10.2020 17:52:32	46.833	6.886	3.4	1.1	4	0.8
62	20.10.2020 18:06:19	46.833	6.886	3.4	1.8	4	0.7
63	20.10.2020 18:36:42	46.833	6.886	3.3	0.7	4	0.7
71	14.12.2020 22:09:25	46.699	6.651	8.8	1.4	7	6.7
87	24.02.2021 18:07:16	46.699	6.652	8.7	1.2	7	6.5
88	07.03.2021 15:31:58	46.846	6.733	2.7	0.1	1	-1.4
89	07.03.2021 15:41:16	46.846	6.733	2.7	0.2	1	-0.6
91	14.03.2021 02:45:38	46.846	6.733	2.7	1.1	1	-0.2
92	14.03.2021 05:17:33	46.846	6.733	2.7	1.5	1	-0.8
93	14.03.2021 05:32:46	46.846	6.733	2.6	0.6	1	1.1
95	18.03.2021 05:07:38	46.846	6.733	2.7	0.5	1	0.5
96	25.03.2021 12:49:41	46.845	6.733	2.6	1.1	1	-0.8
101	21.05.2021 01:52:55	46.925	6.825	42.9	2.5	16	42.9
102	22.05.2021 15:05:28	46.919	6.831	42.8	0.8	16	42.8
103	05.06.2021 17:25:36	46.846	6.732	2.9	1.8	1	2.1
104	06.06.2021 13:21:58	46.789	6.675	1.8	0.6	2	0.5
105	07.06.2021 16:11:22	46.744	6.644	2.2	1.7	10	0.1
115	28.06.2021 04:52:09	46.923	6.829	42.7	0.8	16	42.7
125	17.07.2021 22:45:26	46.743	6.645	2	0.1	10	-0.1

Figure J.5: Relocated events

Appendix K

Template Matching : Templates and detections

K.1 Template creation

Table K.1: Parameters for template creation using EQcorrscan

Channel	Z, N, E
Phases	P, S
Bandpass filter (Hz)	1-30
Pre-pick time (s)	0.1
Sampling rate (sample/s)	125
Length (s)	4
Minimum SNR	5.0
Minimum number of waveforms	4
Processing length (s)	86400

K.2 Templates from LAN stations

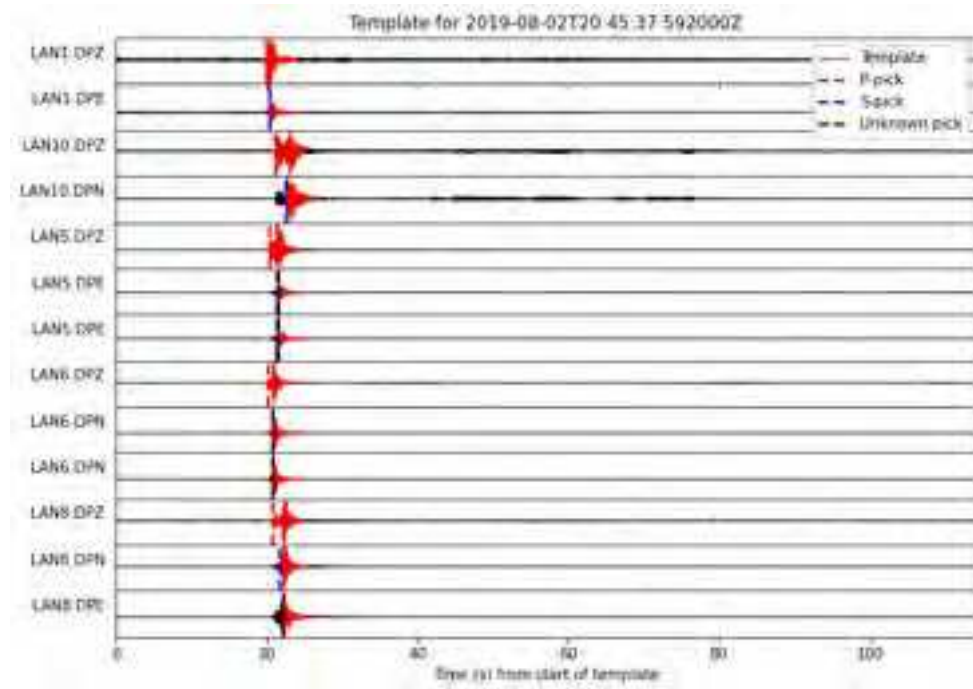


Figure K.1: Event used as template for template matching technique. In red portion of the waveform used as template.

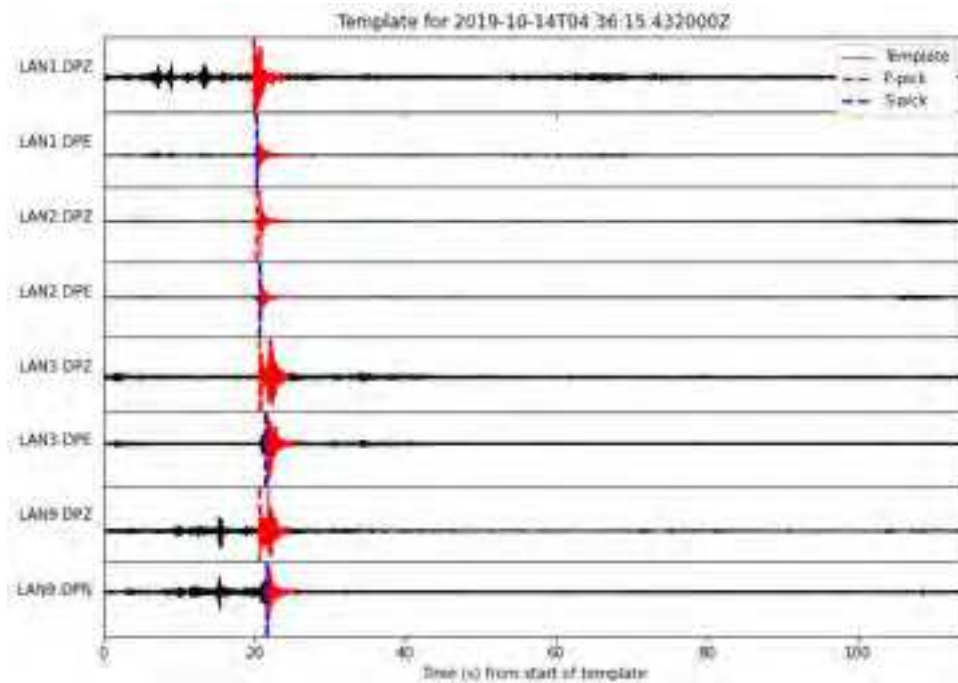


Figure K.2: Event used as template for template matching technique. In red portion of the waveform used as template.

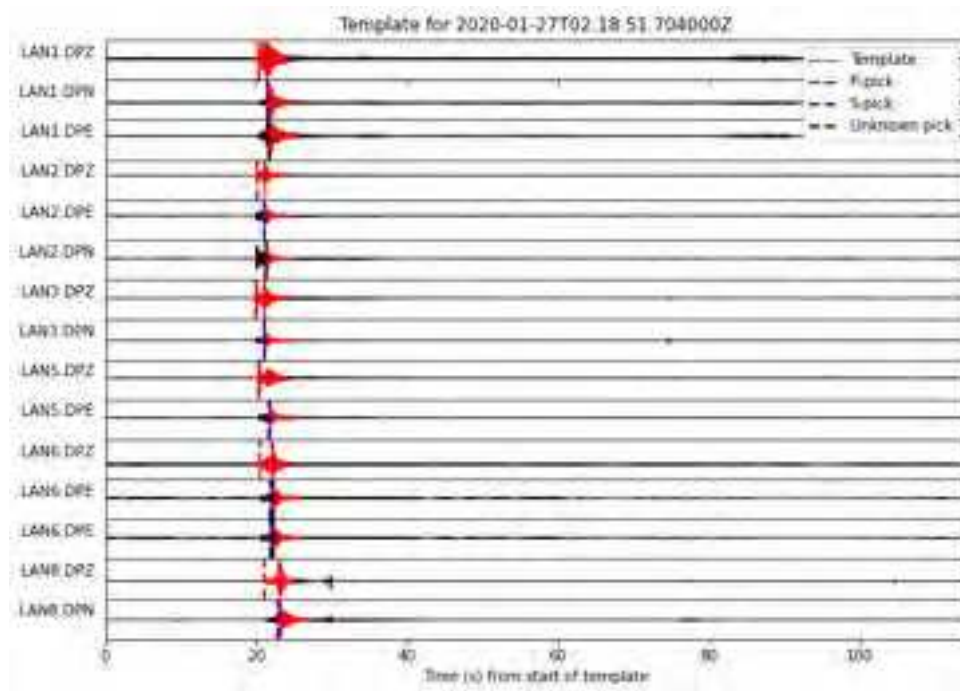


Figure K.3: Event used as template for template matching technique. In red portion of the waveform used as template.

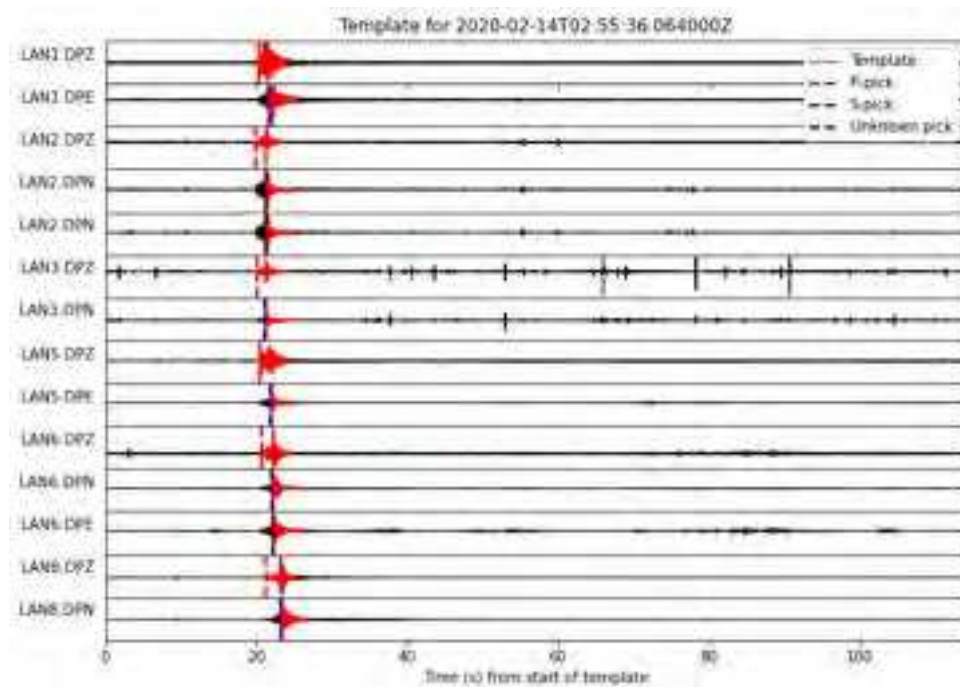


Figure K.4: Event used as template for template matching technique. In red portion of the waveform used as template.

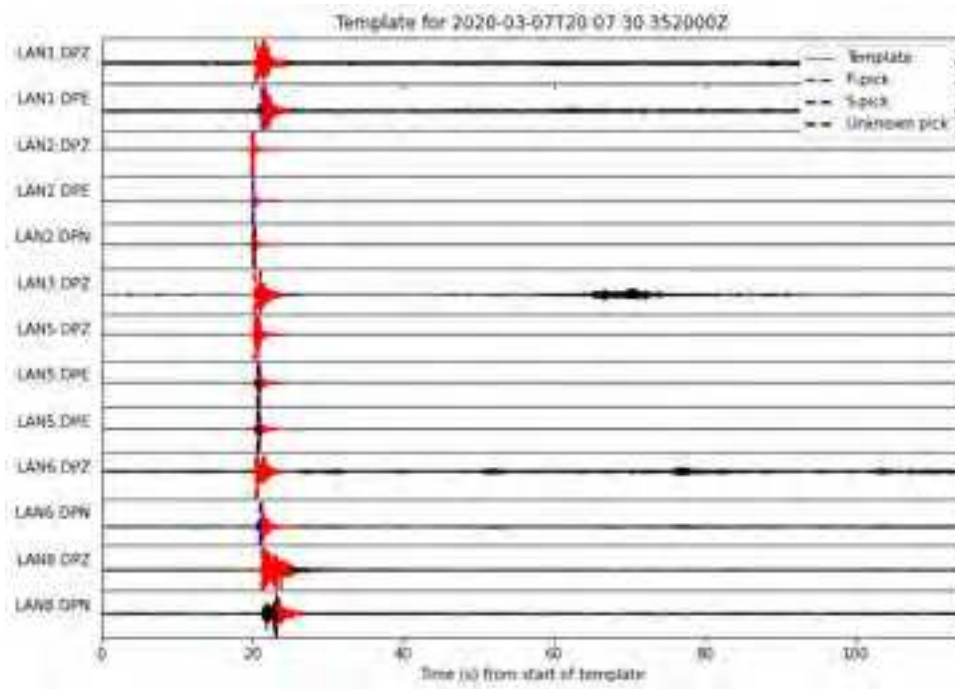


Figure K.5: Event used as template for template matching technique. In red portion of the waveform used as template.

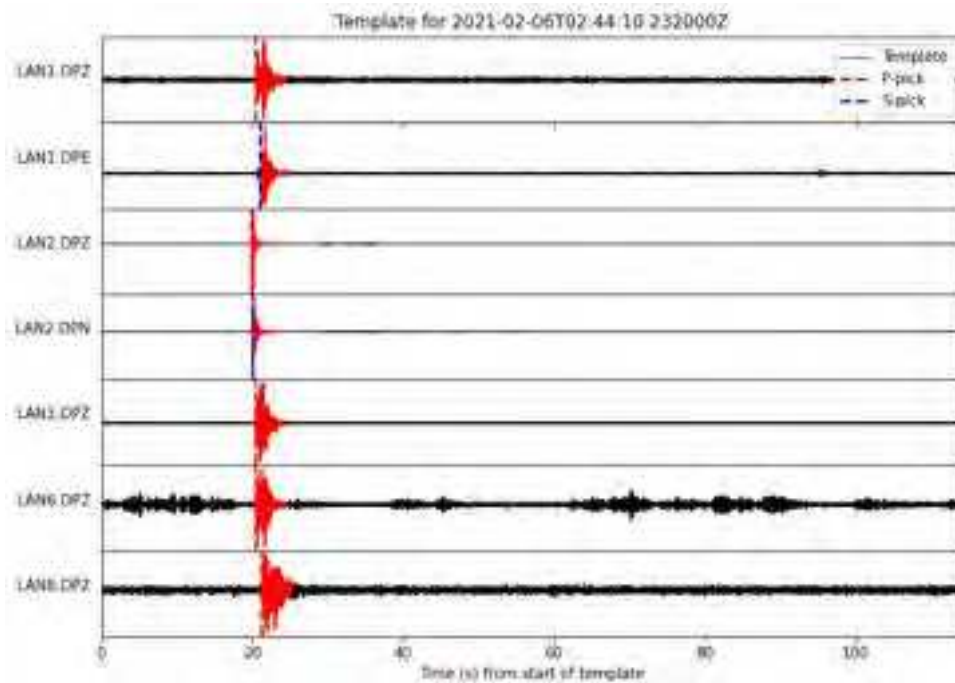


Figure K.6: Event used as template for template matching technique. In red portion of the waveform used as template.

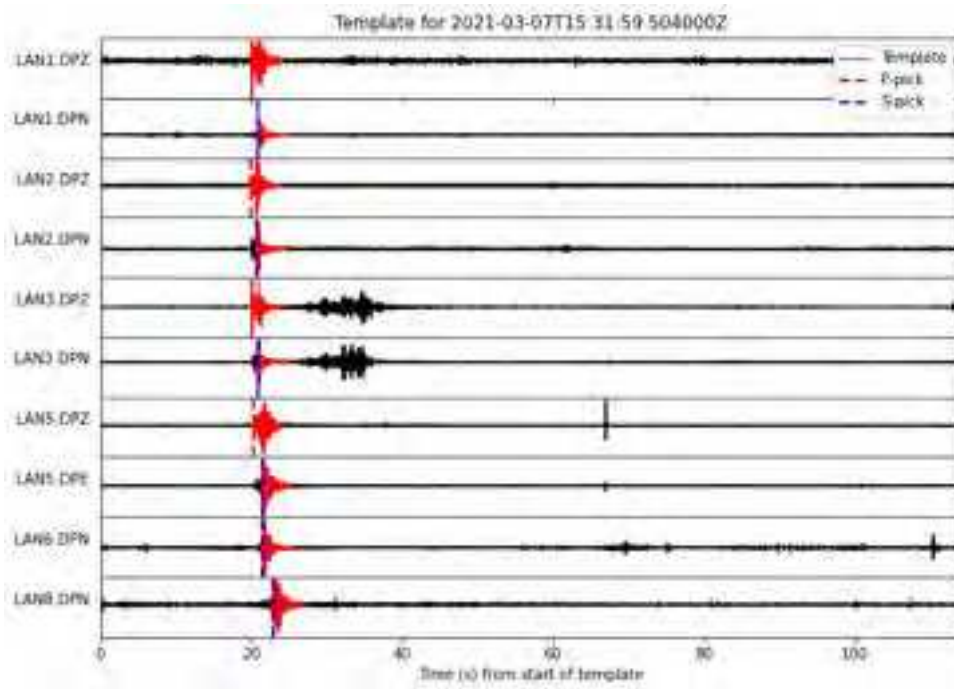


Figure K.7: Event used as template for template matching technique. In red portion of the waveform used as template.

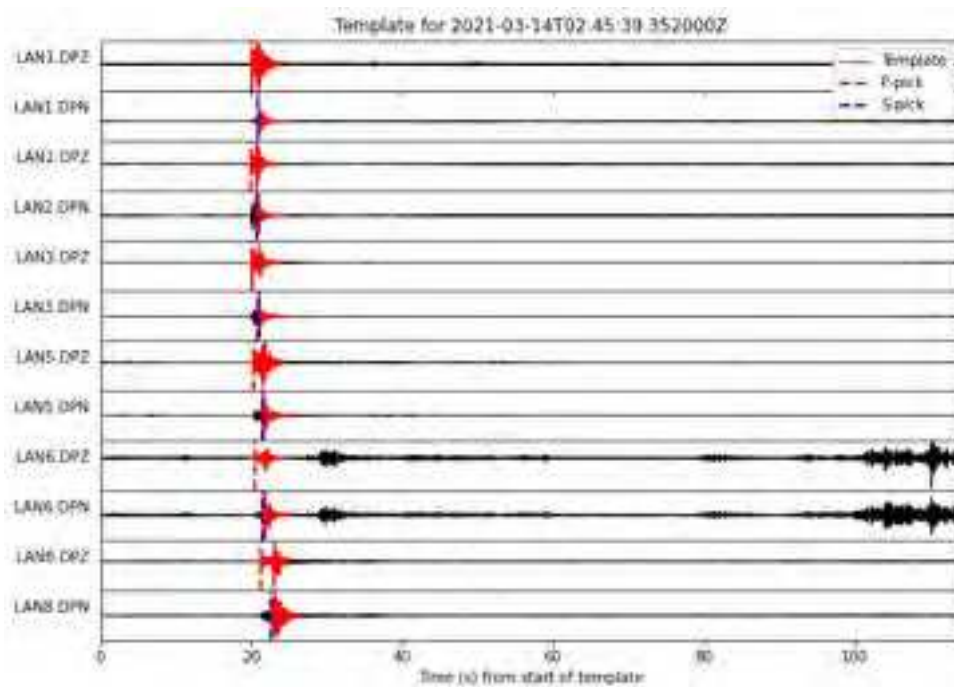


Figure K.8: Event used as template for template matching technique. In red portion of the waveform used as template.

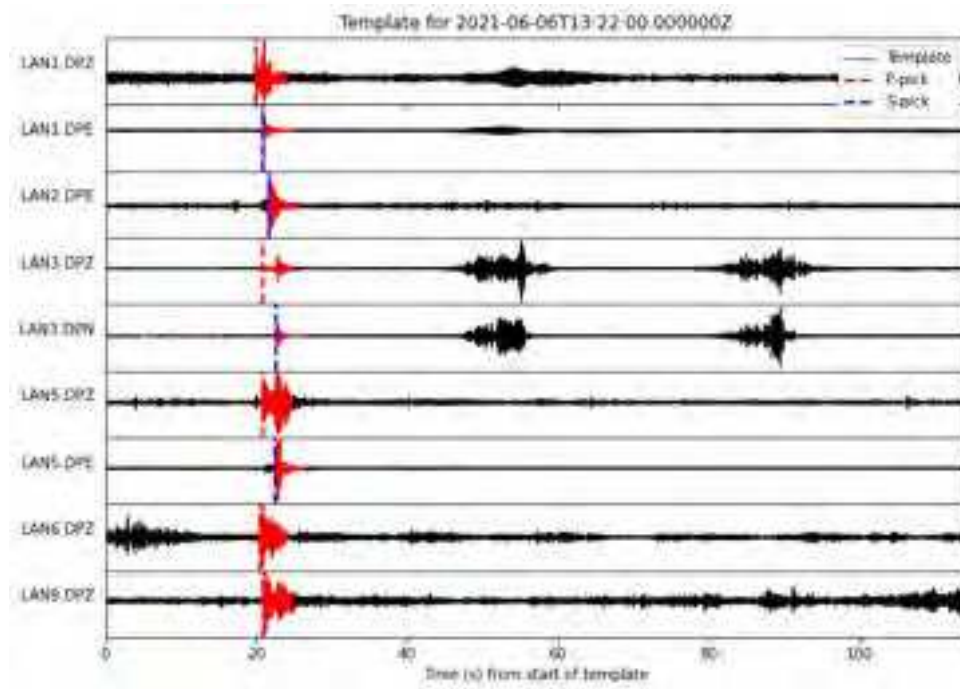


Figure K.9: Event used as template for template matching technique. In red portion of the waveform used as template.

K.3 Templates from EST stations

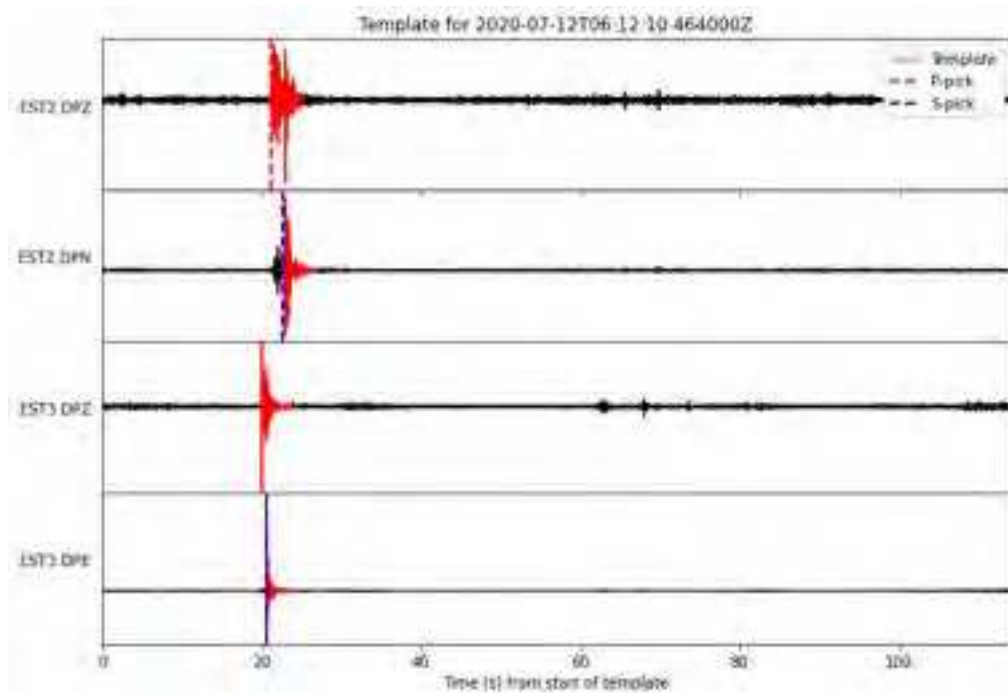


Figure K.10: Event used as template for template matching technique. In red portion of the waveform used as template.

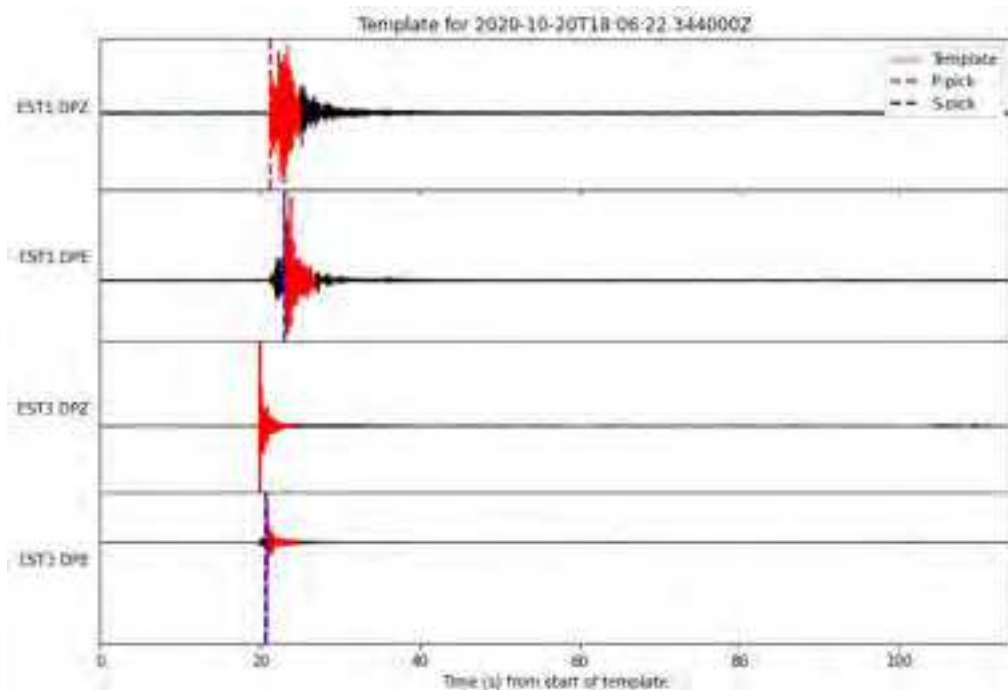


Figure K.11: Event used as template for template matching technique. In red portion of the waveform used as template.

K.4 Templates from BOU stations

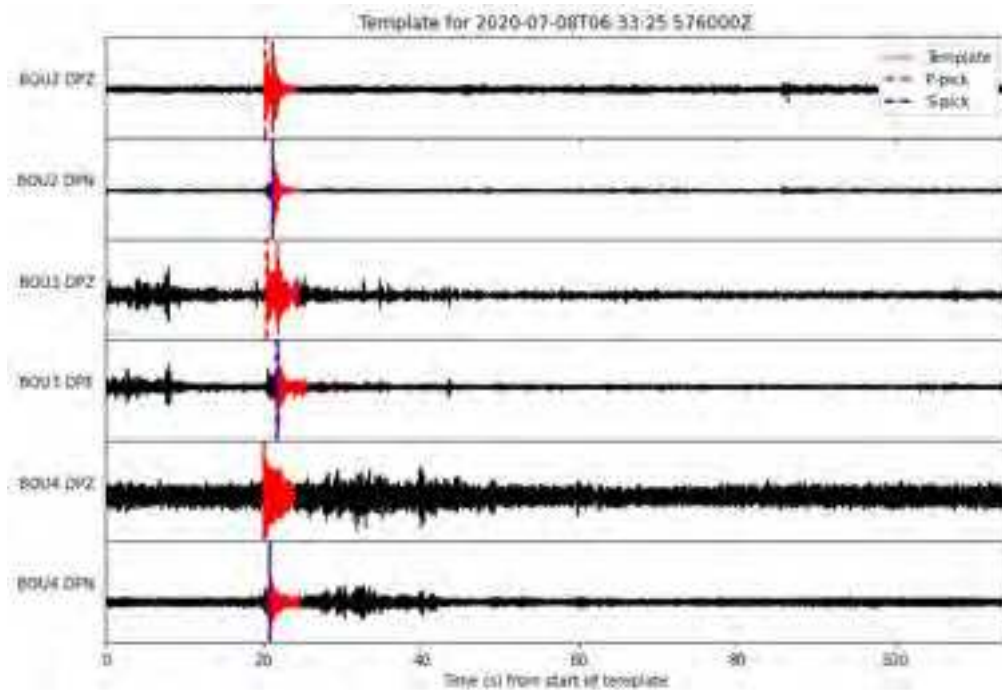


Figure K.12: Event used as template for template matching technique. In red portion of the waveform used as template.

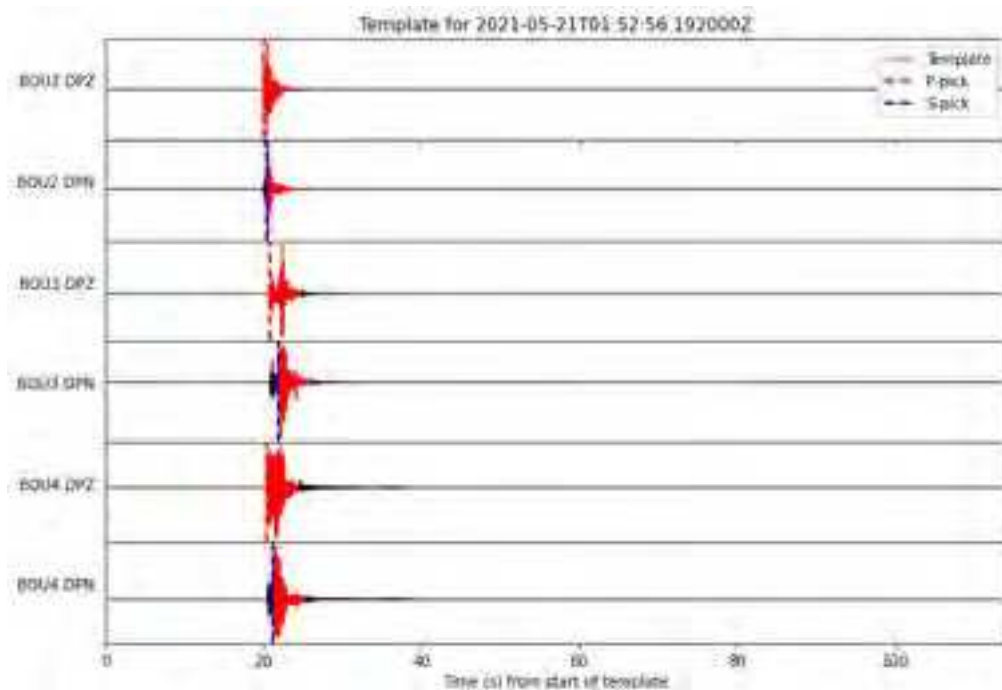


Figure K.13: Event used as template for template matching technique. In red portion of the waveform used as template.

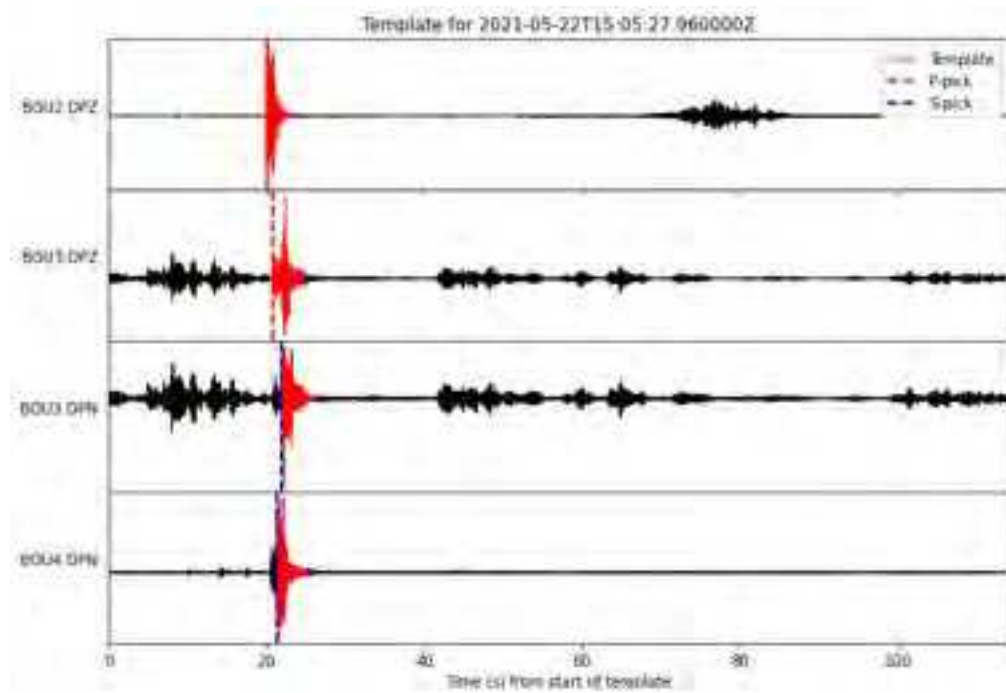


Figure K.14: Event used as template for template matching technique. In red portion of the waveform used as template.

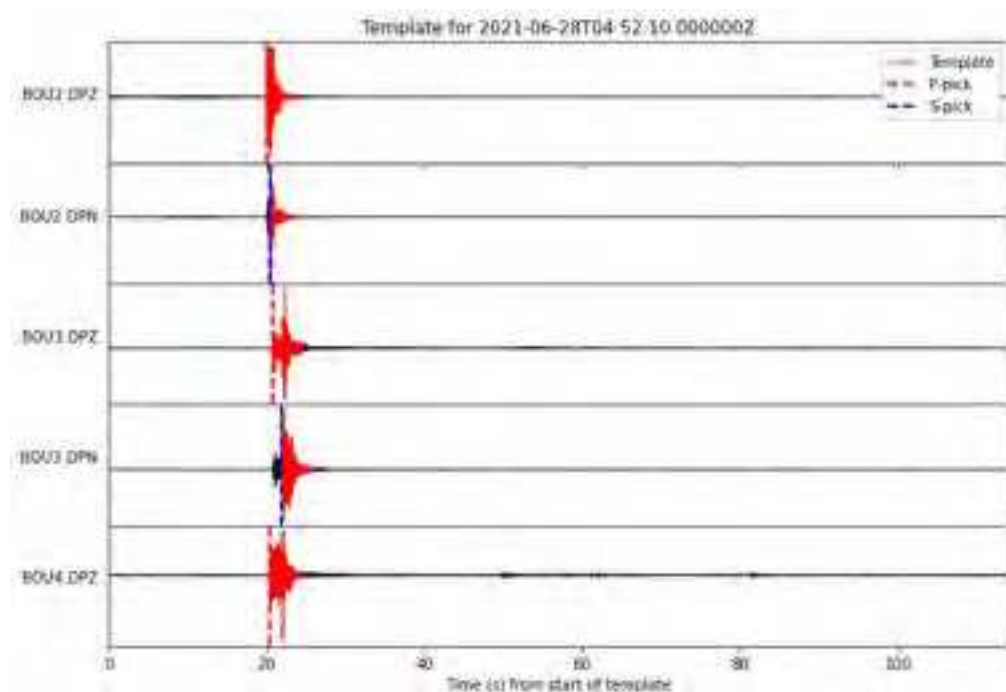


Figure K.15: Event used as template for template matching technique. In red portion of the waveform used as template.

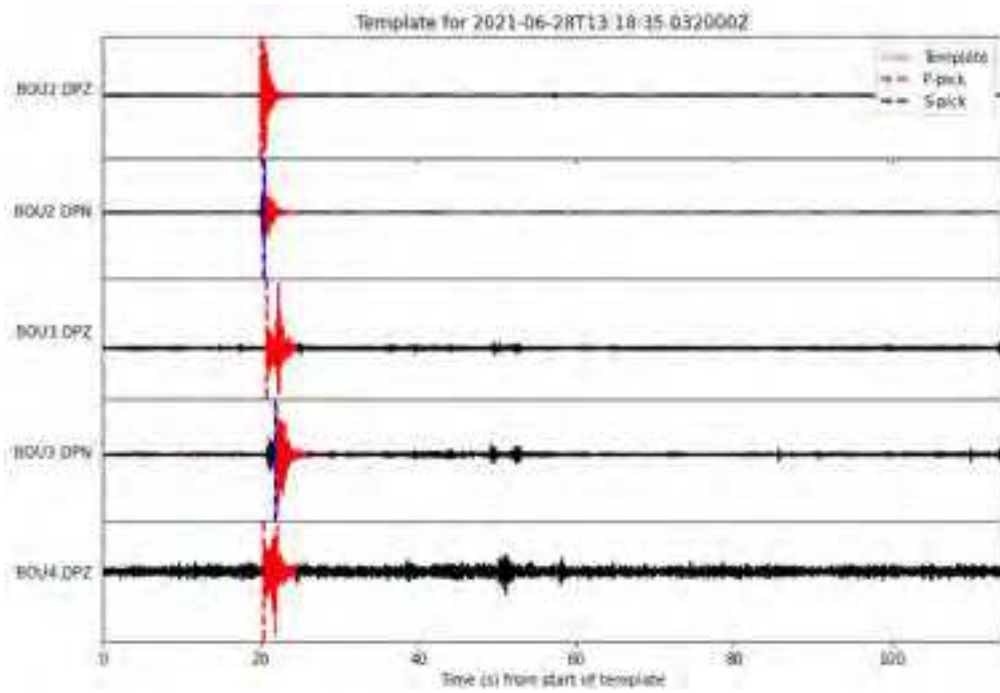


Figure K.16: Event used as template for template matching technique. In red portion of the waveform used as template.

K.5 Templates from FER stations

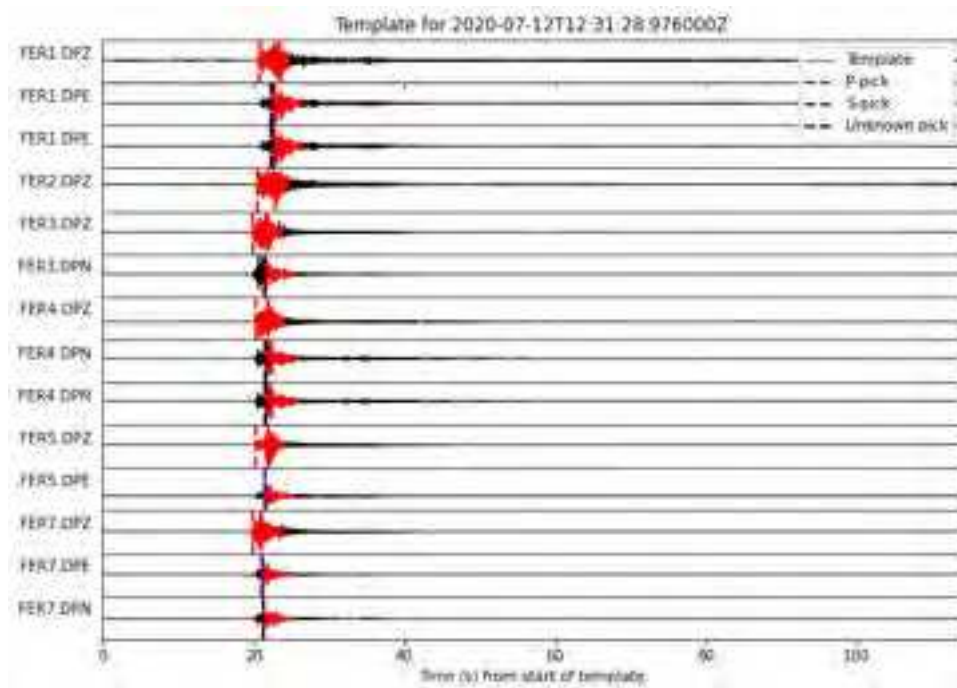


Figure K.17: Event used as template for template matching technique. In red portion of the waveform used as template.

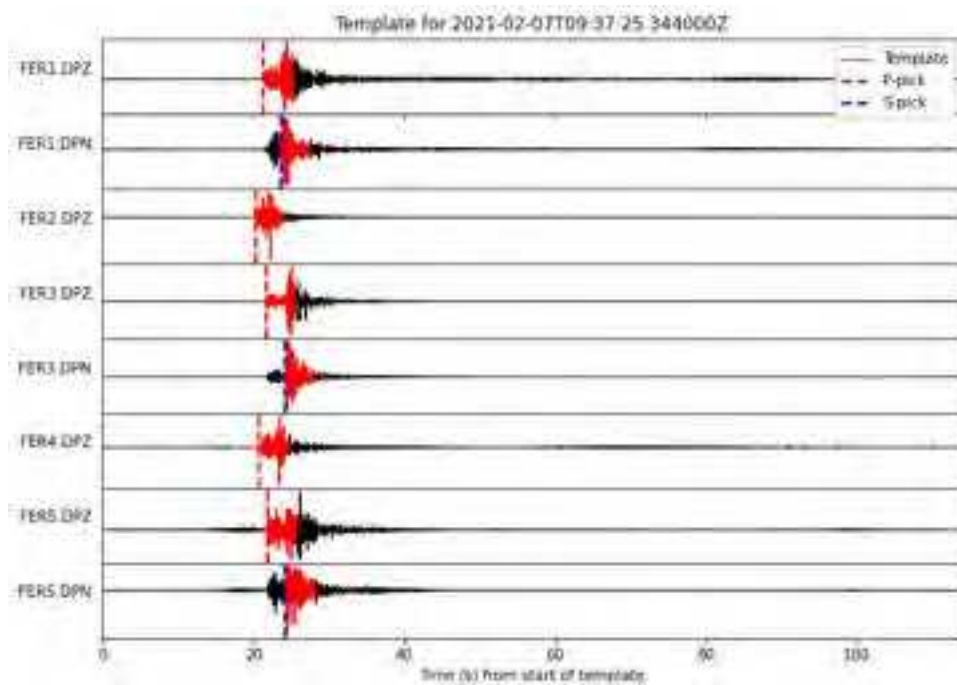


Figure K.18: Event used as template for template matching technique. In red portion of the waveform used as template.

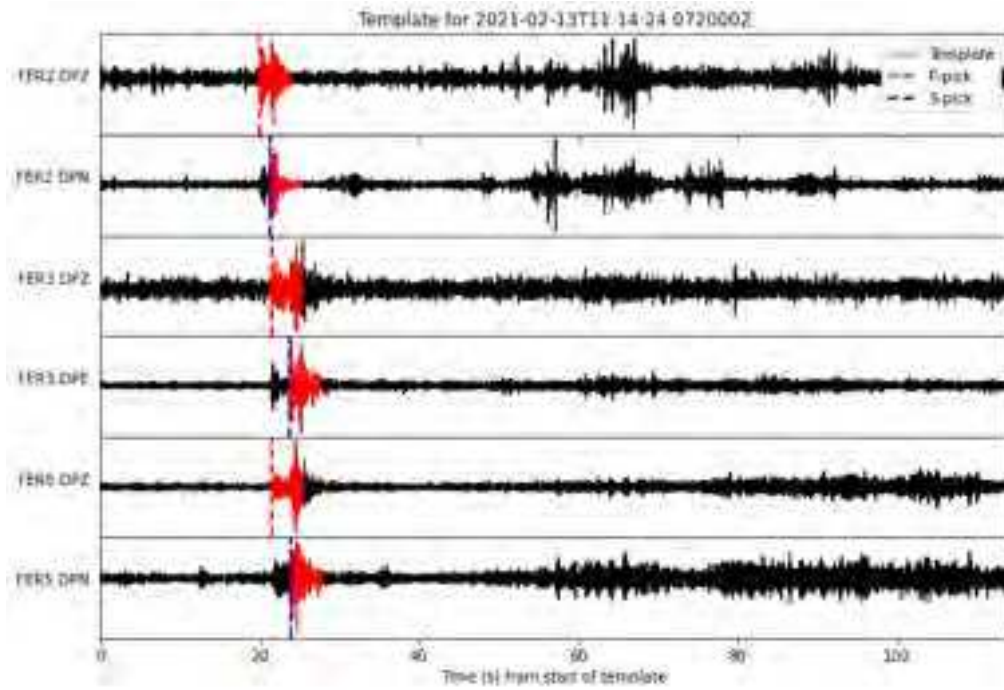


Figure K.19: Event used as template for template matching technique. In red portion of the waveform used as template.

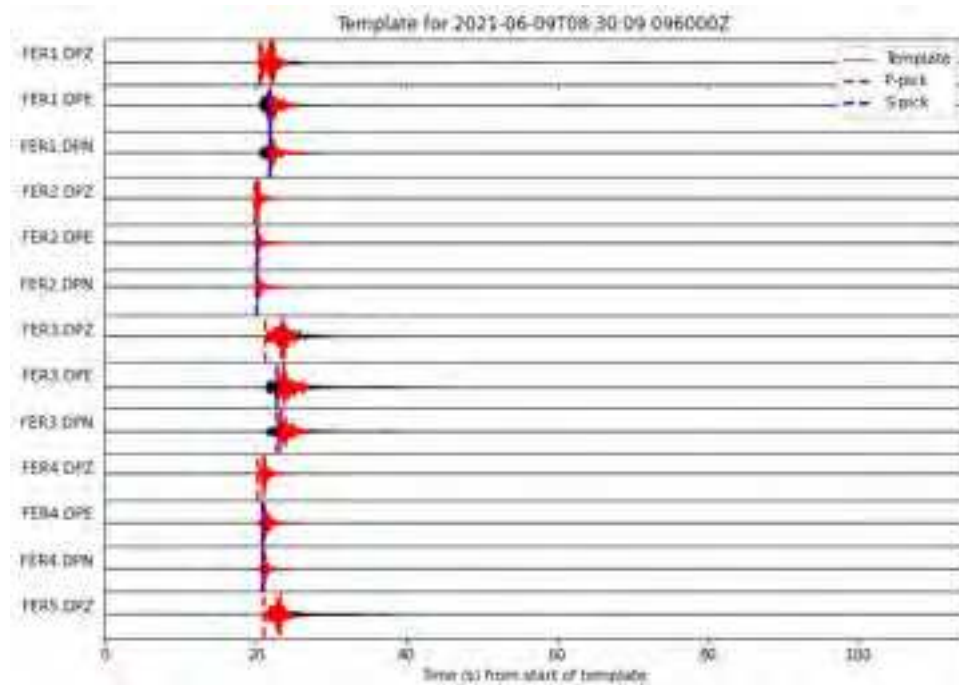


Figure K.20: Event used as template for template matching technique. In red portion of the waveform used as template.

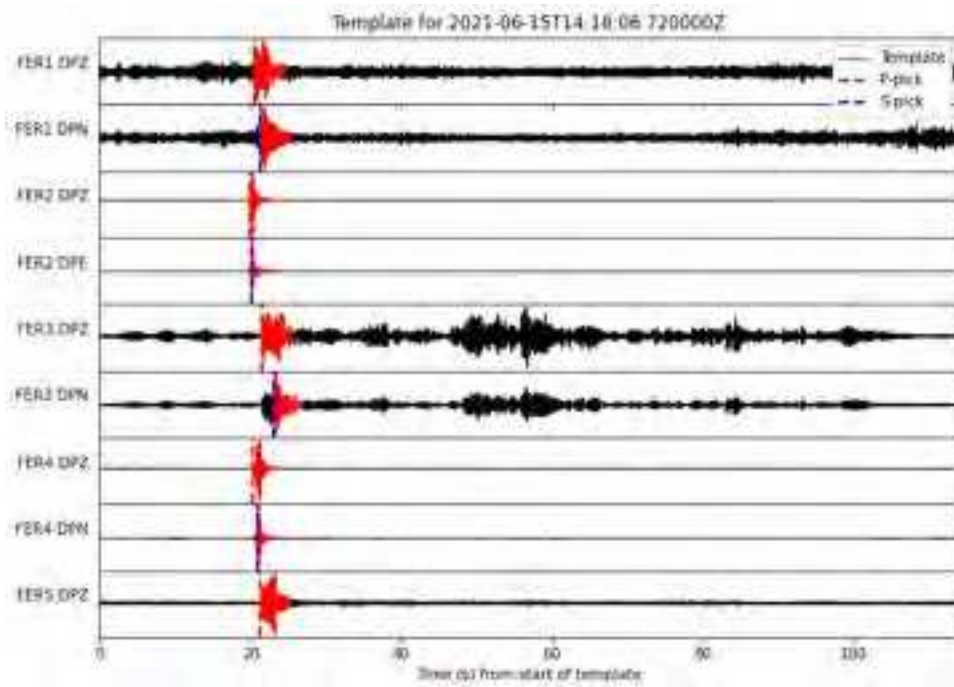


Figure K.21: Event used as template for template matching technique. In red portion of the waveform used as template.

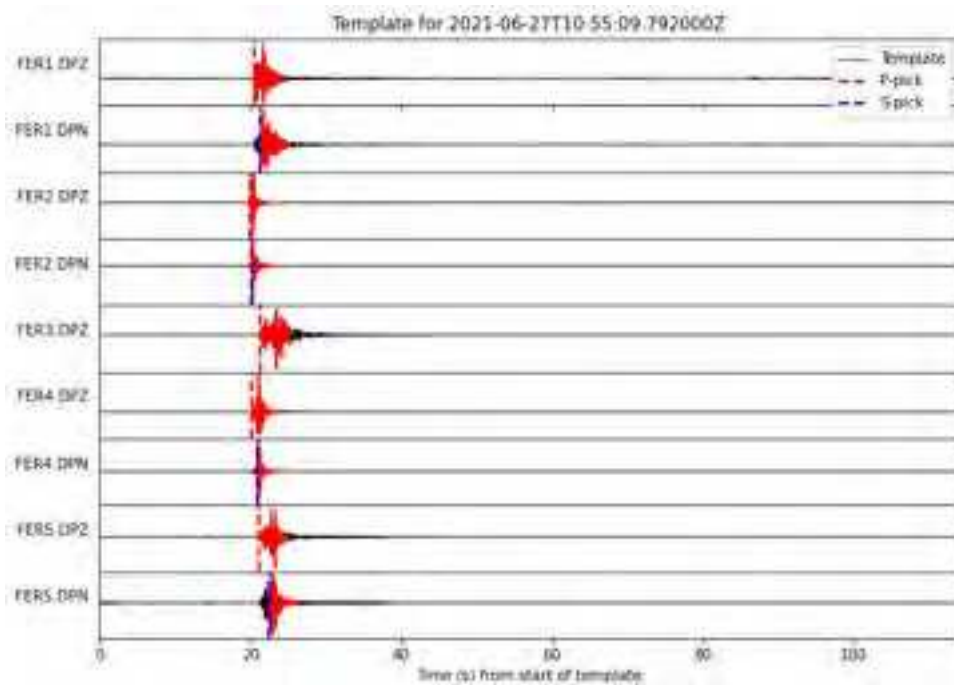
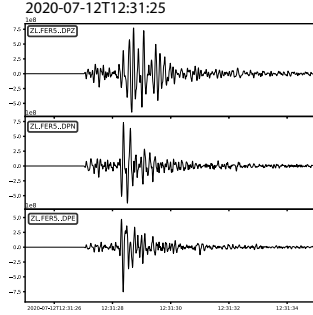


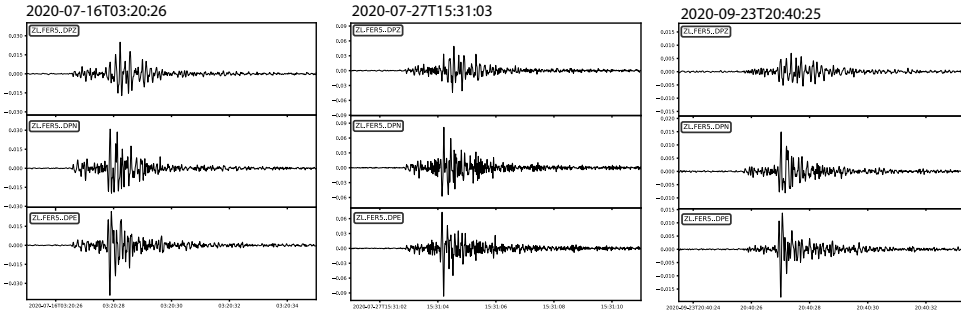
Figure K.22: Event used as template for template matching technique. In red portion of the waveform used as template.

K.6 Example of waveforms similarity between templates and detections

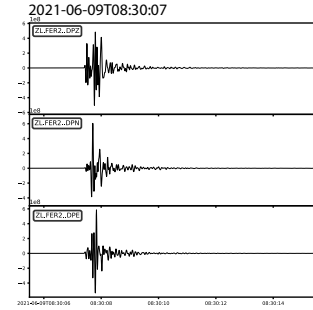
TEMPLATE on station FER5



DETECTIONS on station FER5



TEMPLATE on station FER2



DETECTIONS on station FER2

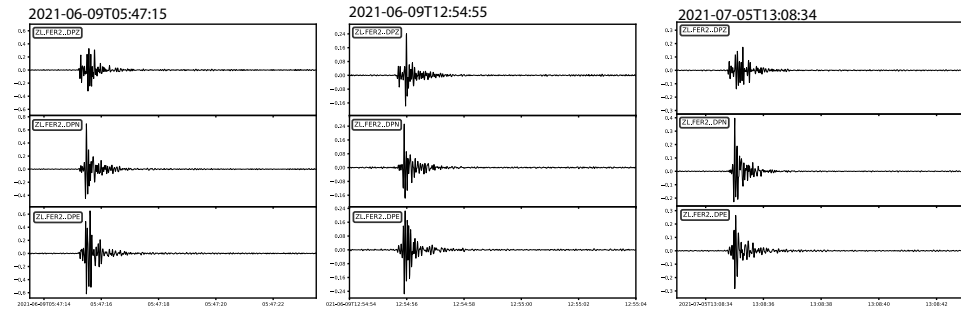


Figure K.23: 2 examples of templates on a single station and their detections on the same station. Top: template on station FER5 for the event 2020-07-12T12:31:25 and detections for this template on the 2020-07-16T03:20:26, 2020-07-27T15:31:03 and 2020-09-23T20:40:25. Bottom: template on station FER2 for the event 2021-06-09T08:30:07 and detections for this template on the 2021-06-09T05:47:15, 2021-06-09T12:54:55 and 2021-07-05T13:08:34.

K.7 List of detections from templates

Figures [K.24](#)[K.25](#) show the detections made from the templates. The detection value CCCS corresponds to the cross-correlation coefficient sum. The sum of the cross-correlation coefficient for every channel is compared to a threshold value. The threshold is set by multiplying the Median Absolute Deviation (MAD) of the CCCS of a day by a factor of 14. This factor was set in order to limit false detections. Max CCCS correspond to the value if the detection was a perfect match.

Template name	Detection time	Detection value (CCCS)	Max CCCS (perfect match)
2019_08_02t20_45_37	2019-07-16T23:31:35.464000Z	4.80	13
2019_08_02t20_45_37	2019-08-03T22:18:25.400000Z	4.58	13
2019_10_14t04_36_15	2019-10-13T00:21:45.896000Z	3.30	8
2019_10_14t04_36_15	2019-10-16T05:05:59.440000Z	3.25	8
2020_01_27t02_18_50	2020-01-16T01:05:33.488000Z	8.97	15
2020_02_14t02_55_34	2020-02-14T21:40:51.712000Z	5.41	14
2020_03_07t20_07_29	2020-03-07T21:05:55.848000Z	6.24	13
2020_03_07t20_07_29	2021-02-02T10:09:37.472000Z	2.64	10
2020_03_07t20_07_29	2021-02-04T04:49:06.328000Z	2.66	10
2020_03_07t20_07_29	2021-02-04T18:14:58.864000Z	2.76	10
2020_03_07t20_07_29	2021-02-06T08:57:05.784000Z	3.56	10
2020_07_08t06_33_24	2020-07-05T02:34:58.624000Z	4.72	6
2020_07_12t06_12_07	2020-05-20T17:32:26.112000Z	2.37	4
2020_07_12t06_12_07	2020-05-23T12:59:47.048000Z	3.13	4
2020_07_12t06_12_07	2020-05-23T14:46:10.744000Z	2.50	4
2020_07_12t06_12_07	2020-05-23T14:54:14.856000Z	1.67	4
2020_07_12t06_12_07	2020-05-27T23:05:06.880000Z	1.59	4
2020_07_12t06_12_07	2020-06-13T03:38:10.992000Z	1.67	4
2020_07_12t06_12_07	2020-07-06T10:10:54.832000Z	3.69	4
2020_07_12t06_12_07	2020-07-12T21:08:03.544000Z	2.48	4
2020_07_12t06_12_07	2020-07-24T00:46:37.776000Z	1.94	4
2020_07_12t12_31_26	2020-07-11T05:42:42.544000Z	6.10	14
2020_07_12t12_31_26	2020-07-12T19:18:28.592000Z	4.19	14
2020_07_12t12_31_26	2020-07-12T20:40:23.936000Z	3.40	14
2020_07_12t12_31_26	2020-07-16T03:20:26.192000Z	10.92	14
2020_07_12t12_31_26	2020-07-27T15:31:02.528000Z	9.56	14
2020_07_12t12_31_26	2020-08-21T05:00:32.848000Z	5.47	14
2020_07_12t12_31_26	2020-09-23T20:40:25.328000Z	6.82	14
2020_10_20t18_06_19	2020-07-07T11:16:04.448000Z	1.76	4
2020_10_20t18_06_19	2020-07-07T21:16:23.320000Z	1.77	4
2020_10_20t18_06_19	2020-10-15T01:00:07.904000Z	2.88	4
2020_10_20t18_06_19	2020-10-15T01:00:25.264000Z	2.28	4
2020_10_20t18_06_19	2020-10-17T22:16:28.600000Z	2.08	4
2020_10_20t18_06_19	2020-10-17T22:18:01.624000Z	1.55	4
2020_10_20t18_06_19	2020-10-20T16:08:35.816000Z	1.77	4
2020_10_20t18_06_19	2020-10-20T16:14:08.416000Z	2.82	4
2020_10_20t18_06_19	2020-10-20T16:43:54.904000Z	2.22	4
2020_10_20t18_06_19	2020-10-20T17:36:22.048000Z	2.02	4
2020_10_20t18_06_19	2020-10-20T17:39:26.520000Z	2.89	4
2020_10_20t18_06_19	2020-10-20T17:47:28.192000Z	2.54	4
2020_10_20t18_06_19	2020-10-20T17:47:43.968000Z	2.14	4
2020_10_20t18_06_19	2020-10-20T18:04:16.528000Z	2.39	4
2020_10_20t18_06_19	2020-10-20T18:07:25.688000Z	1.74	4
2020_10_20t18_06_19	2020-10-20T18:09:28.256000Z	2.71	4
2020_10_20t18_06_19	2020-10-20T18:10:56.208000Z	1.60	4
2020_10_20t18_06_19	2020-10-20T18:17:35.000000Z	2.72	4
2020_10_20t18_06_19	2020-10-20T18:24:28.520000Z	1.69	4
2020_10_20t18_06_19	2020-10-20T18:35:55.304000Z	2.05	4
2020_10_20t18_06_19	2020-10-20T18:56:50.624000Z	1.61	4
2020_10_20t18_06_19	2020-10-20T18:57:51.400000Z	2.96	4
2020_10_20t18_06_19	2020-10-20T19:21:52.040000Z	3.29	4
2020_10_20t18_06_19	2020-10-20T19:22:42.584000Z	2.25	4
2020_10_20t18_06_19	2020-10-20T20:34:24.648000Z	2.20	4
2020_10_20t18_06_19	2020-10-20T20:53:03.616000Z	1.92	4
2020_10_20t18_06_19	2020-10-20T21:48:24.864000Z	3.43	4
2021_02_06t02_44_09	2021-02-06T02:44:18.904000Z	2.39	7
2021_02_07t09_37_22	2021-02-18T03:56:41.776000Z	2.93	8
2021_02_13t11_14_22	2021-02-11T22:20:36.352000Z	3.36	6
2021_02_13t11_14_22	2021-02-11T22:21:39.256000Z	3.54	6
2021_02_13t11_14_22	2021-02-11T22:22:02.072000Z	4.99	6

Figure K.24: List of detections from templates. Page 1/2

2021_02_13t11_14_22	2021-02-11T22:45:44.496000Z	5.09	6
2021_02_13t11_14_22	2021-02-11T22:46:37.672000Z	3.81	6
2021_03_07t15_31_58	2021-03-07T15:33:31.672000Z	6.14	10
2021_03_07t15_31_58	2021-03-07T15:35:24.864000Z	4.59	10
2021_03_07t15_31_58	2021-03-07T15:35:55.240000Z	3.37	10
2021_03_07t15_31_58	2021-03-07T16:16:32.192000Z	4.84	10
2021_03_14t02_45_38	2021-03-07T15:24:48.672000Z	8.97	12
2021_03_14t02_45_38	2021-03-07T15:25:03.648000Z	6.68	12
2021_03_14t02_45_38	2021-03-07T15:25:54.864000Z	6.74	12
2021_03_14t02_45_38	2021-03-07T15:29:46.464000Z	6.86	12
2021_03_14t02_45_38	2021-03-07T15:33:54.584000Z	5.94	12
2021_03_14t02_45_38	2021-03-07T15:41:01.664000Z	7.06	12
2021_03_14t02_45_38	2021-03-07T15:41:49.568000Z	5.58	12
2021_03_14t02_45_38	2021-03-08T01:46:32.040000Z	6.11	12
2021_03_14t02_45_38	2021-03-14T02:47:35.480000Z	9.08	12
2021_03_14t02_45_38	2021-03-19T02:48:58.112000Z	3.20	12
2021_03_14t02_45_38	2021-06-02T16:08:49.840000Z	7.57	12
2021_03_14t02_45_38	2021-06-07T00:09:37.616000Z	5.46	12
2021_05_21t01_52_55	2021-05-21T02:27:02.208000Z	3.02	6
2021_05_21t01_52_55	2021-05-21T03:21:35.800000Z	1.75	6
2021_05_21t01_52_55	2021-05-21T06:43:27.528000Z	1.73	6
2021_05_21t01_52_55	2021-05-21T14:39:09.096000Z	2.89	6
2021_05_21t01_52_55	2021-05-21T16:49:50.016000Z	3.08	6
2021_05_21t01_52_55	2021-06-01T22:14:11.064000Z	2.88	6
2021_05_21t01_52_55	2021-07-24T01:19:17.792000Z	1.67	6
2021_05_21t01_52_55	2021-07-24T15:26:06.280000Z	1.69	6
2021_05_22t15_05_28	2021-05-21T01:56:00.352000Z	2.25	4
2021_06_06t13_21_59	2021-06-07T01:08:33.056000Z	4.96	9
2021_06_09t08_30_07	2021-06-09T05:47:15.152000Z	8.36	13
2021_06_09t08_30_07	2021-06-09T05:47:26.144000Z	4.04	13
2021_06_09t08_30_07	2021-06-09T12:54:55.552000Z	3.71	13
2021_06_09t08_30_07	2021-06-10T14:28:21.848000Z	4.26	13
2021_06_09t08_30_07	2021-06-15T10:49:27.000000Z	3.61	13
2021_06_09t08_30_07	2021-07-05T13:08:34.664000Z	5.78	13
2021_06_15t14_18_05	2021-06-09T08:42:47.272000Z	3.46	9
2021_06_15t14_18_05	2021-06-15T14:20:25.720000Z	3.62	9
2021_06_15t14_18_05	2021-06-15T21:52:47.352000Z	3.26	9
2021_06_15t14_18_05	2021-06-17T06:59:41.072000Z	6.06	9
2021_06_15t14_18_05	2021-06-27T11:18:37.472000Z	4.39	9
2021_06_15t14_18_05	2021-06-27T11:22:47.392000Z	6.53	9
2021_06_27t10_55_08	2021-06-28T22:14:50.232000Z	5.35	9
2021_06_28t04_52_09	2021-05-21T03:58:53.752000Z	1.73	5
2021_06_28t04_52_09	2021-06-28T06:24:10.592000Z	3.03	5
2021_06_28t04_52_09	2021-07-19T17:00:13.744000Z	1.68	5
2021_06_28t13_18_34	2021-05-21T01:55:05.192000Z	1.92	5
2021_06_28t13_18_34	2021-06-28T05:14:58.136000Z	2.89	5

Figure K.25: List of detections from templates. Page 2/2

Appendix L

Chi-squared statistical test

Figure L.1 shows the observed and expected values used for the chi-squared statistical test.

Null hypothesis : There is no link between high groundwater levels in the Torrent sinkhole and the occurrence of seismicity

Water level [m]	Hours		Tot2
	Without EQ	With EQ	
≥ 100	816 824	14 6	830
< 100	15750 15742	103 111	15853
Tot1	16566	117	Total number of hours = 16683

Observed values

Expected value = $(\text{Tot1} * \text{Tot2}) / \text{Tot hours}$

Results

$$\chi^2 = \sum \frac{(\text{Observed} - \text{Expected})^2}{\text{Expected}}$$

Chi-squared table					
P	0.25	0.1	0.05	0.005	0.0005
χ^2	1.323	2.706	3.841	7.879	12.116

Limit of level of significance:

$\chi^2 = 12.116 \rightarrow$ P-value close to 0.0005 for threshold of 100 m

Figure L.1: Chi-squared statistical test for the 100 m threshold, stating the Null hypothesis, showing the observed (green) and expected (red) number of earthquakes occurring during hours of high groundwater levels (≥ 100 m) and low groundwater levels (< 100 m) and the resulting value in comparison to the chi-squared table for 1 degree of freedom. The limit of significance is marked by a red line. If the result is higher than 7.879, the null hypothesis can be rejected.

Appendix M

Analytical modelling: Results

M.1 Pore pressure diffusion

Pore pressure response for every scenario of hydraulic diffusivity and distance is shown for every earthquake considered as wet (Figures [M.1](#)[M.2](#)). The first column is the ID of the event and the second column is the origin time of the event. Columns 3-6 are pore pressure values resulting from different hydraulic diffusivity. Distances, from 100 to 5000 m, are indicated in front of the pore pressure values. The underlined values correspond to plausible scenarios, i.e., rupture occurred during an increasing pore pressure or a punctual maximum.

ID	origin time	D = 0.1 m ² /s	D = 1 m ² /s	D = 10 m ² /s	D = 100 m ² /s
EQ-1	2019-10-21 02:44:28	<u>100 0.444679</u>	100 0.621659	100 0.667610	100 0.681324
		1000 0.127788	<u>1000 0.096532</u>	<u>1000 0.444679</u>	1000 0.621659
		2000 0.162809	2000 0.089843	<u>2000 0.219057</u>	2000 0.543768
		3500 0.131488	3500 0.136203	<u>3500 0.083049</u>	<u>3500 0.416296</u>
		5000 0.106371	5000 0.159061	5000 0.075444	<u>5000 0.300906</u>
EQ-2	2020-02-05 23:07:04	100 0.711802	100 0.826788	100 0.858201	100 0.867919
		1000 0.185190	<u>1000 0.331962</u>	1000 0.711802	1000 0.826788
		2000 0.162828	<u>2000 0.172031</u>	<u>2000 0.516717</u>	2000 0.777032
		3500 0.131447	3500 0.187947	<u>3500 0.293279</u>	3500 0.691547
		5000 0.108730	5000 0.179359	<u>5000 0.194656</u>	<u>5000 0.598165</u>
C13-1	2020-03-07 20:07:29	<u>100 0.824810</u>	100 0.964682	100 0.999687	100 1.010068
		1000 0.249888	<u>1000 0.477478</u>	<u>1000 0.824810</u>	1000 0.964682
		2000 0.167638	<u>2000 0.324843</u>	<u>2000 0.625849</u>	2000 0.904171
		3500 0.132069	3500 0.235237	3500 0.449487	3500 0.801479
		5000 0.109089	5000 0.189382	<u>5000 0.369047</u>	<u>5000 0.702536</u>
C13-2	2021-02-05 14:12:09	<u>100 0.946587</u>	<u>100 1.099047</u>	100 1.145212	100 1.159660
		1000 0.114480	<u>1000 0.509343</u>	<u>1000 0.946587</u>	<u>1000 1.099047</u>
		2000 0.120571	2000 0.187290	<u>2000 0.727149</u>	2000 1.029468
		3500 0.125834	3500 0.112715	<u>3500 0.456285</u>	<u>3500 0.922458</u>
		5000 0.113509	5000 0.117134	<u>5000 0.277841</u>	<u>5000 0.816752</u>
C17-1	2021-02-07 09:37:19	100 0.947278	100 1.076795	100 1.117519	100 1.130383
		1000 0.122308	<u>1000 0.564048</u>	1000 0.947278	1000 1.076795
		2000 0.120511	<u>2000 0.226898</u>	<u>2000 0.760674</u>	2000 1.017015
		3500 0.125684	3500 0.116640	<u>3500 0.513376</u>	<u>3500 0.927025</u>
		5000 0.113461	5000 0.116994	<u>5000 0.331000</u>	<u>5000 0.837729</u>
C9-1	2021-02-10 16:05:09	100 0.905614	100 0.961676	100 0.974059	100 0.977737
		1000 0.142993	<u>1000 0.623219</u>	1000 0.905614	1000 0.961676
		2000 0.120387	<u>2000 0.291101</u>	<u>2000 0.782850</u>	2000 0.939434
		3500 0.125423	3500 0.129166	<u>3500 0.578545</u>	3500 0.894296
		5000 0.113373	5000 0.117323	<u>5000 0.404702</u>	5000 0.837995
C1-1	2021-03-07 15:31:58	100 0.171555	100 0.063451	100 0.022322	100 0.009025
		1000 0.249393	1000 0.311707	1000 0.171555	1000 0.063451
		2000 0.131618	2000 0.318761	2000 0.263794	2000 0.118407
		3500 0.123656	3500 0.228192	3500 0.318985	3500 0.184788
		5000 0.112659	5000 0.158053	5000 0.329706	5000 0.233028
C16-1	2021-05-21 01:52:55	100 0.589480	100 0.633426	100 0.637976	100 0.638840
		<u>1000 0.156518</u>	<u>1000 0.362617</u>	1000 0.589480	1000 0.633426
		2000 0.155442	<u>2000 0.194691</u>	<u>2000 0.484135</u>	2000 0.618360
		3500 0.125740	3500 0.155797	<u>3500 0.333123</u>	<u>3500 0.579575</u>
		5000 0.110672	5000 0.159731	<u>5000 0.239189</u>	<u>5000 0.530713</u>

Figure M.1: Pore pressure response from diffusion process at time of earthquakes. Page 1/2

C5-1	2021-06-30 23:49:02	100 0.573197	100 0.722836	100 0.770465	100 0.785512
		1000 0.179058	1000 0.254566	1000 0.573197	1000 0.722836
		2000 0.157592	2000 0.177798	2000 0.384925	2000 0.653126
		3500 0.128098	3500 0.179394	3500 0.232928	3500 0.550574
		5000 0.110354	5000 0.170250	5000 0.187350	5000 0.456333
EQ-3	2021-07-14 01:38:01	100 0.680762	100 0.972617	100 1.082203	100 1.117488
		1000 0.198905	1000 0.353621	1000 0.680762	1000 0.972617
		2000 0.158278	2000 0.250910	2000 0.453940	2000 0.824421
		3500 0.128638	3500 0.189975	3500 0.336147	3500 0.645647
		5000 0.110370	5000 0.169088	5000 0.281822	5000 0.523901
C16-2	2021-07-19 17:00:13	100 0.891287	100 0.963070	100 0.981683	100 0.987402
		1000 0.211840	1000 0.587273	1000 0.891287	1000 0.963070
		2000 0.158601	2000 0.310089	2000 0.752142	2000 0.932819
		3500 0.128883	3500 0.199636	3500 0.544529	3500 0.877929
		5000 0.110388	5000 0.171095	5000 0.393916	5000 0.813287
C17-2	2021-07-21 03:01:40	100 0.811872	100 0.830963	100 0.831151	100 0.830960
		1000 0.217356	1000 0.605591	1000 0.811872	1000 0.830963
		2000 0.158755	2000 0.332792	2000 0.732307	2000 0.826268
		3500 0.128945	3500 0.203198	3500 0.568338	3500 0.805901
		5000 0.110394	5000 0.171801	5000 0.423413	5000 0.771285

Figure M.2: HypoDD output file. Page 2/2

Appendix N

Flash flood in Cressier

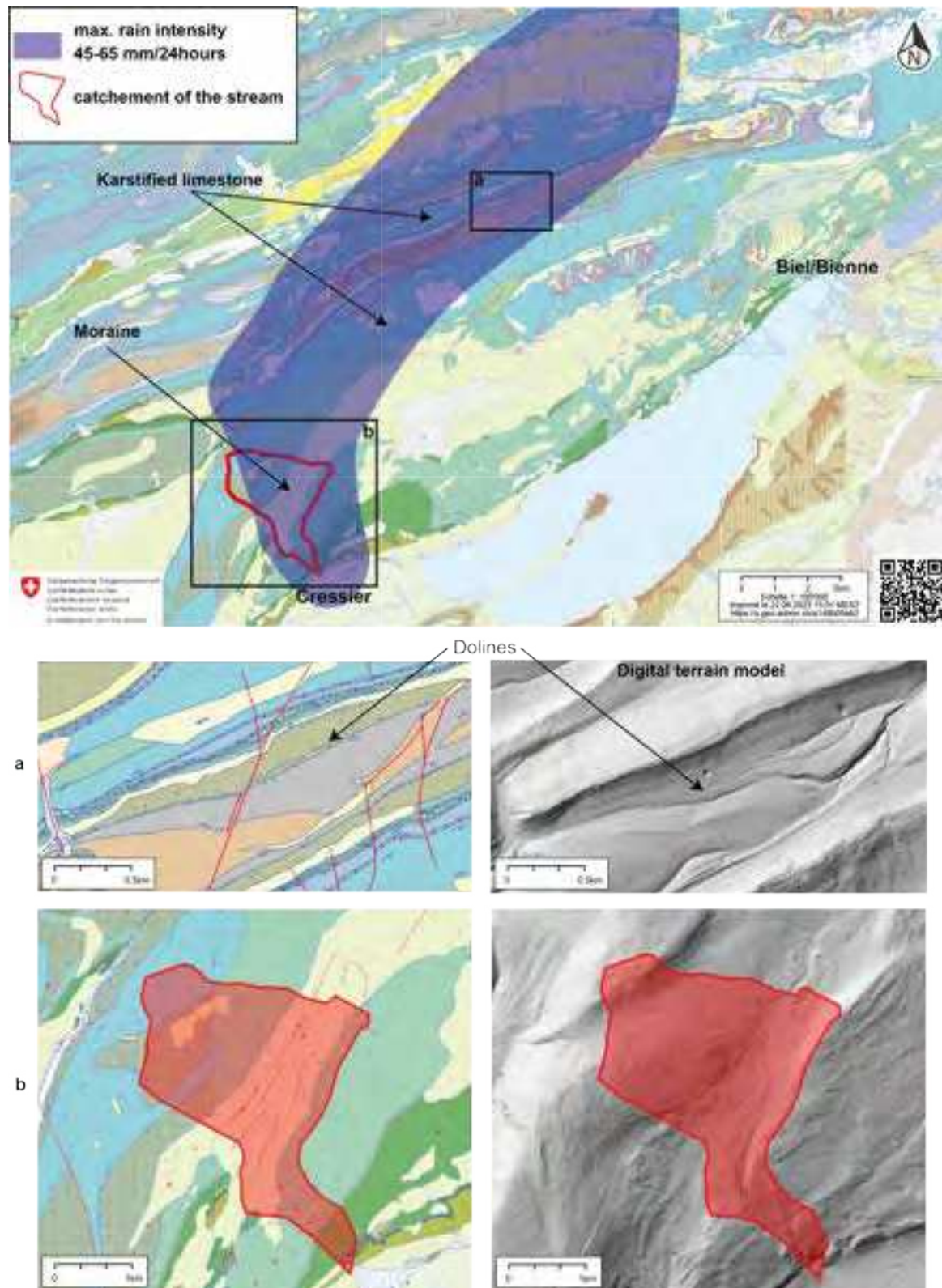


Figure N.1: Geological map and location of the rainfall event of June 22 2021. Most of the rain fell on highly fractured and karstified limestones as attested by the numerous dolines reported on the geological map and visible on the digital terrain model (a). The catchment area of the stream is shown in red. This area shows no reported dolines and almost no fault segments, suggesting that runoff might be important, partly explaining the flash flood that occurred in June 2021.

Bibliography

- Abednego, M., 2015: Microseismic Tomography Analysis of the larger Fribourg area (western Swiss Molasse Basin). *GeoFocus*, **40**, 162 pp.
- Anderson, E. M., 1951: *The Dynamics of Faulting and Dyke Formation with Applications to Britain*. 2nd ed., Oliver & Boyd, Edimburgh.
- Angelier, J., 1979: Determination of the mean principal directions of stresses for a given fault population. *Tectonophysics*, **56** (3-4), T17–T26.
- Angelier, J., 2002: Inversion of earthquake focal mechanisms to obtain the seismotectonic stress IV-a new method free of choice among nodal planes. *Geophys. J. Int.*, **150**, 588–609, URL <https://academic.oup.com/gji/article/150/3/588/613206>.
- Antunes, V., T. Planes, J. Zahradník, A. Obermann, C. Alvizuri, A. Carrier, and M. Lupi, 2020: Seismotectonics and 1-D velocity model of the Greater Geneva Basin, France-Switzerland. *Geophysical Journal International*, **221** (3), 2026–2047, doi:10.1093/GJI/GGAA129.
- Atkinson, T. C., 1977: Diffuse flow and conduit flow in limestone terrain in the Mendip Hills, Somerset (Great Britain). *Journal of Hydrology*, **35**, 93.
- Audra, P., L. Mocochain, H. Camus, E. Gilli, G. Clauzon, and J. Y. Bigot, 2004: The effect of the Messinian Deep Dstage on karst development around the Mediterranean Sea. Examples from Southern France. *Geodinamica Acta*, **17** (6), 389–400, doi:10.3166/ga.17.389-400.
- Audra, P., and A. N. Palmer, 2013: The Vertical Dimension of Karst: Controls of Vertical Cave Pattern. *Treatise on Geomorphology*, Vol. 6, Elsevier Inc., 186–206, doi:10.1016/B978-0-12-374739-6.00098-1.
- Baer, M., and Coauthors, 2007: Earthquakes in Switzerland and surrounding regions during 2006. *Swiss Journal of Geosciences*, **100** (3), 517–528, doi:10.1007/s00015-007-1242-0.
- Barton, C. A., M. D. Zoback, and D. Moos, 1995: Fluid flow along potentially active faults in crystalline rock. *Geology*, **23** (8), 683–686, doi:10.1130/0091-7613(1995)023<0683:FFAPAF>2.3.CO;2.
- Barton, N., 1973: Review of a new shear-strength criterion for rock joints. *Engineering Geology*, **7** (4), 287–332, doi:10.1016/0013-7952(73)90013-6.

- Bayle, B., and D. Graillot, 1987: Fontaine de Vaucluse : Compte-rendu hydrogéologique de l'opération Spélénaute du 2 août 1985. *Karstologia : revue de karstologie et de spéléologie physique*, **9** (1), 1–6, doi:10.3406/karst.1987.2150.
- Becker, A., 2000: The Jura Mountains - an active foreland fold-and-thrust belt? *Tectonophysics*, **321**, 381–406.
- Becker, A., and D. Werner, 1995: Etat des contraintes neotectoniques dans le jura. *Geodinamica Acta*, **8** (2), 99–111, doi:10.1080/09853111.1995.11105277.
- Beeler, N. M., R. W. Simpson, S. H. Hickman, and D. A. Lockner, 2000: Pore fluid pressure, apparent friction, and Coulomb failure. *Journal of Geophysical Research: Solid Earth*, **105** (B11), 25 533–25 542, doi:10.1029/2000jb900119.
- Bettinelli, P., J.-P. Avouac, M. Flouzat, L. Bollinger, G. Ramillien, S. Rajaure, and S. Sapkota, 2008: Seasonal variations of seismicity and geodetic strain in the Himalaya induced by surface hydrology. *Earth and Planetary Science Letters*, **266** (3), 332–344, doi:https://doi.org/10.1016/j.epsl.2007.11.021, URL <http://www.sciencedirect.com/science/article/pii/S0012821X07007492>.
- Beyreuther, M., R. Barsch, L. Krischer, T. Megies, Y. Behr, and J. Wassermann, 2010: ObsPy: A Python Toolbox for Seismology. *Seismological Research Letters*, **81**, 530–533, doi:10.1785/gssrl.81.3.530.
- Birhanu, Y., M. Wilks, J. Biggs, J. M. Kendall, A. Ayele, and E. Lewi, 2018: Seasonal patterns of seismicity and deformation at the Alutu geothermal reservoir, Ethiopia, induced by hydrological loading. *Journal of Volcanology and Geothermal Research*, **356**, 175–182, doi:10.1016/j.jvolgeores.2018.03.008.
- Bollinger, L., F. Perrier, J. P. Avouac, S. Sapkota, U. Gautam, and D. R. Tiwari, 2007: Seasonal modulation of seismicity in the Himalaya of Nepal. *Geophysical Research Letters*, **34** (8), doi:10.1029/2006GL029192.
- Bonacci, O., 1987: *Karst hydrology : with special reference to the Dinaric karst*. 1st ed., Springer Series in Physical Environment, 2, Springer-Verlag, Berlin.
- Bonacci, O., 1993: Karst springs hydrographs as indicators of karst aquifers. *Hydrological Sciences Journal*, **38** (1), 51–62, doi:10.1080/02626669309492639.
- Bott, M. H., 1959: The Mechanics of Oblique Slip Faulting. *Geological Magazine*, **96** (2), 109–117, doi:10.1017/S0016756800059987.
- Brace, W. F., and J. D. Byerlee, 1966: Stick-Slip as a Mechanism for Earthquakes. *Science*, **153**, 990–992, doi:10.1126/science.153.3739.990.
- Bragato, P. L., 2021: Systematic Triggering of Large Earthquakes by Karst Water Recharge: Statistical Evidence in Northeastern Italy. *Frontiers in Earth Science*, **9** (July), 1–11, doi:10.3389/feart.2021.664932.

- Braunmiller, J., J. L. Nábělek, and A. M. Tréhu, 2014: A seasonally modulated earthquake swarm near Maupin, Oregon. *Geophysical Journal International*, **197** (3), 1736–1743, doi:10.1093/gji/ggu081.
- Burkhard, M., 1993: Calcite twins, their geometry, appearance and significance as stress-strain markers and indicators of tectonic regime: a review. Tech. rep., 351 pp.
- Burkhard, M., and A. Sommaruga, 1998: Evolution of the western Swiss Molasse basin: structural relations with the Alps and the Jura belt. *Geological Society Special Publication*, **134**, 279–298, doi:10.1144/GSL.SP.1998.134.01.13.
- Byerlee, J., 1978: Friction of rocks. *Pure and Applied Geophysics PAGEOPH*, **116** (4-5), 615–626, doi:10.1007/BF00876528.
- Caine, J. S., J. P. Evans, and C. B. Forster, 1996: Fault zone architecture and permeability structure. *Geology*, **24** (11), 1025–1028.
- Carslaw, H. S., and J. Jaeger, 1959: *Conduction of heat in solids*. Second edition. Oxford : Clarendon Press, 1959., 510 pp., URL <https://search.library.wisc.edu/catalog/999607365802121>.
- Chamberlain, C. J., C. J. Hopp, C. M. Boese, E. Warren-Smith, D. Chambers, S. X. Chu, K. Michailos, and J. Townend, 2018: EQcorrscan: Repeating and near-repeating earthquake detection and analysis in python. *Seismological Research Letters*, **89** (1), 173–181, doi:10.1785/0220170151.
- Charmoille, A., O. Fabbri, J. Mudry, Y. Guglielmi, and C. Bertrand, 2005: Post-seismic permeability change in a shallow fractured aquifer following a ML 5.1 earthquake (Fourbanne karst aquifer, Jura outermost thrust unit, eastern France). *Geophysical Research Letters*, **32** (18), 1–5, doi:10.1029/2005GL023859.
- Clauzon, G., 1982: The Messinian Rhone canyon as a definite proof of the "desiccated deep-basin model". *Bulletin de la Societe Geologique de France*, **S7-XXIV** (3), 597–610, doi:10.2113/gssgfbull.S7-XXIV.3.597.
- Cornet, F. H., 2016: Seismic and aseismic motions generated by fluid injections. *Geomechanics for Energy and the Environment*, **5**, 42–54, doi:10.1016/j.gete.2015.12.003.
- Costain, J. K., and G. A. Bollinger, 2010: Review: Research results in hydroseismicity from 1987 to 2009. *Bulletin of the Seismological Society of America*, **100** (5 A), 1841–1858, doi:10.1785/0120090288.
- Crank, J., 1975: *The Mathematics of Diffusion*. 2nd ed., Clarendon Press, Oxford.
- Diehl, T., E. Kissling, and P. Bormann, 2012: Tutorial for consistent phase picking at local to regional distances. *New Manual of Seismological Observatory Practice (NMSOP-2)*, P. Bormann, Ed., GFZ German Research Centre for Geosciences, Potsdam, 1–21, doi:10.2312/GFZ.NMSOP-2{_}IS{_}11.4.

- Diehl, T., T. Kraft, E. Kissling, and S. Wiemer, 2017: The induced earthquake sequence related to the St. Gallen deep geothermal project (Switzerland): Fault reactivation and fluid interactions imaged by microseismicity. *Journal of Geophysical Research: Solid Earth*, **122** (9), 7272–7290, doi:10.1002/2017JB014473.
- Dieterich, J. H., 1979: Modeling of rock friction 1. Experimental results and constitutive equations. *Journal of Geophysical Research: Solid Earth*, **84** (B5), 2161–2168, doi:10.1029/JB084iB05p02161.
- Duboeuf, L., L. De Barros, F. Cappa, Y. Guglielmi, A. Deschamps, and S. Seguy, 2017: Aseismic Motions Drive a Sparse Seismicity During Fluid Injections Into a Fractured Zone in a Carbonate Reservoir. *Journal of Geophysical Research: Solid Earth*, **122** (10), 8285–8304, doi:10.1002/2017JB014535.
- Edwards, B., T. Kraft, C. Cauzzi, P. Kästli, and S. Wiemer, 2015: Seismic monitoring and analysis of deep geothermal projects in st Gallen and Basel, Switzerland. *Geophysical Journal International*, **201** (2), 1022–1039, doi:10.1093/gji/ggv059.
- Elkhoury, J. E., E. E. Brodsky, and D. C. Agnew, 2006: Seismic waves increase permeability. *Nature*, **441** (7097), 1135–1138, doi:10.1038/nature04798.
- Ellsworth, W. L., 2013: Injection-Induced Earthquakes. *Science*, **341** (July), 1–8.
- Evans, J. P., C. B. Forster, and J. V. Goddard, 1997: Permeability of fault-related rocks, and implications for hydraulic structure of fault zones. *Journal of Structural Geology*, **19** (11), 1393–1404, doi:10.1016/S0191-8141(97)00057-6.
- Evans, K. F., 2005: Permeability creation and damage due to massive fluid injections into granite at 3.5 km at Soultz: 2. Critical stress and fracture strength. *Journal of Geophysical Research: Solid Earth*, **110** (4), 1–14, doi:10.1029/2004JB003169.
- Evans, K. F., R. O. Burford, and G. C. P. King, 1981: Propagating episodic creep and the aseismic slip behavior of the Calaveras Fault north of Hollister, California. *Journal of Geophysical Research*, **86** (B5), 3721, doi:10.1029/jb086ib05p03721.
- Evans, K. F., A. Zappone, T. Kraft, N. Deichmann, and F. Moia, 2012: A survey of the induced seismic responses to fluid injection in geothermal and CO₂ reservoirs in Europe. *Geothermics*, **41**, 30–54, doi:10.1016/j.geothermics.2011.08.002, URL <http://dx.doi.org/10.1016/j.geothermics.2011.08.002>.
- Fäh, D., D. Giardini, and F. Bay, 2003: Earthquake Catalogue of Switzerland (ECOS) and the related macroseismic database. *Eclogae Geologicae Helvetiae*, **96** (2), 219–236, doi:10.5169/seals-169017, URL <http://doi.org/10.5169/seals-169017>.
- Faulkner, D. R., C. A. Jackson, R. J. Lunn, R. W. Schlische, Z. K. Shipton, C. A. Wibberley, and M. O. Withjack, 2010: A review of recent developments concerning the structure, mechanics and fluid flow properties of fault zones. *Journal of Structural Geology*, **32** (11), 1557–1575, doi:10.1016/j.jsg.2010.06.009, URL <http://dx.doi.org/10.1016/j.jsg.2010.06.009>.

- Fetter, C. W., 2001: *Applied hydrogeology*. 4th ed., Pearson Education.
- Ford, D., and P. Williams, 2013: The Karst Rocks. *Karst Hydrogeology and Geomorphology*, John Wiley & Sons, Ltd, chap. 2, 9–38, doi:10.1002/9781118684986.ch2, URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/9781118684986.ch2>.
- Foulger, G. R., M. P. Wilson, J. G. Gluyas, B. R. Julian, and R. J. Davies, 2018: Global review of human-induced earthquakes. *Earth-Science Reviews*, **178** (January 2017), 438–514, doi:10.1016/j.earscirev.2017.07.008, URL <https://doi.org/10.1016/j.earscirev.2017.07.008>.
- Gargani, J., 2004: Modelling of the erosion in the Rhone valley during the Messinian crisis (France). *Quaternary International*, **121** (1), 13–22, doi:10.1016/j.quaint.2004.01.020.
- Gephart, J. W., and D. W. Forsyth, 1984: An improved method for determining the regional stress tensor using earthquake focal mechanism data: application to the San Fernando earthquake sequence. *Journal of Geophysical Research: Solid Earth*, **89** (B11), 9305–9320.
- González, P. J., K. F. Tiampo, M. Palano, F. Cannavó, and J. Fernández, 2012: The 2011 Lorca earthquake slip distribution controlled by groundwater crustal unloading. *Nature Geoscience*, **5** (11), 821–825, doi:10.1038/ngeo1610.
- Gorin, G., D. Morend, and A. Pugin, 2003: Bedrock, Quaternary sediment and recent fault activity in central Lake Neuchâtel, as derived from high-resolution reflection seismics. *Ecologiae Geologicae Helveticae*, **96** (SPEC), doi:10.1007/978-3-0348-7992-7{_}2.
- Gratier, J. P., 2011: Fault permeability and strength evolution related to fracturing and healing episodic processes (year to millennia): the role of pressure solution. *Oil and Gas Science and Technology*, **66** (3), 491–506, doi:10.2516/ogst/2010014.
- Grigoli, F., and Coauthors, 2017: Current challenges in monitoring, discrimination, and management of induced seismicity related to underground industrial activities: A European perspective. *Reviews of Geophysics*, **55** (2), 310–340, doi:10.1002/2016RG000542.
- Groshong, R. H., 1972: Strain calculated from twinning in calcite. *Bulletin of the Geological Society of America*, **83** (7), 2025–2038, doi:10.1130/0016-7606(1972)83[2025:SCFTIC]2.0.CO;2.
- Gruber, M., 2017: Structural investigations of the Western Swiss Molasse Basin - from 2D seismic interpretation to a 3D geological model. Ph.D. thesis, University of Fribourg, 190 pp., Fribourg.
- Guglielmetti, L., and Coauthors, 2013: GeoNE: an integrated project for the exploration of low enthalpy deep aquifers in the canton of Neuchâtel, western Switzerland. *European Geothermal Congress 2013*, 12, June.
- Guglielmi, Y., D. Elsworth, F. Cappa, P. Henry, C. Gout, P. Dick, and J. Durand, 2015: In situ observations on the coupling between hydraulic diffusivity and displacements during fault reactivation in shales. *Journal of Geophysical Research: Solid Earth*, **120** (11), 7729–7748, doi:10.1002/2015JB012158.

- Gunatilake, T., and S. A. Miller, 2022: Spatio-Temporal Complexity of Aftershocks in the Apennines Controlled by Permeability Dynamics and Decarbonization. *Journal of Geophysical Research: Solid Earth*, **127** (6), doi:10.1029/2022JB024154.
- Gupta, H. K., 2002: A review of recent studies of triggered earthquakes by artificial water reservoirs with special emphasis on earthquakes in Koyna, India. *Earth-Science Reviews*, **58** (3), 279–310, doi:https://doi.org/10.1016/S0012-8252(02)00063-6.
- Haimson, B., and F. Cornet, 2003: ISRM Suggested Methods for rock stress estimation—Part 3: hydraulic fracturing (HF) and/or hydraulic testing of pre-existing fractures (HTPF). *International Journal of Rock Mechanics and Mining Sciences*, **40** (7-8), 1011–1020, doi:10.1016/j.ijrmms.2003.08.002.
- Hainzl, S., Y. Ben-Zion, C. Cattania, and J. Wassermann, 2013: Testing atmospheric and tidal earthquake triggering at Mt. Hochstaufen, Germany. *Journal of Geophysical Research: Solid Earth*, **118** (10), 5442–5452, doi:10.1002/jgrb.50387.
- Hainzl, S., T. Kraft, J. Wassermann, H. Igel, and E. Schmedes, 2006: Evidence for rainfall-triggered earthquake activity. *Geophysical Research Letters*, **33** (19), L19303, doi:10.1029/2006GL027642.
- Hardebeck, J. L., and P. M. Shearer, 2002: A New Method for Determining First-Motion Focal Mechanisms. *Bulletin of the Seismological Society of America*, **92** (6), 2264–2276.
- Häring, M. O., U. Schanz, F. Ladner, and B. C. Dyer, 2008: Characterisation of the Basel 1 enhanced geothermal system. *Geothermics*, **37** (5), 469–495, doi:10.1016/j.geothermics.2008.06.002.
- Havskov, J., and L. Ottemöller, 1999: SeisAn Earthquake Analysis Software. *Seismological Research Letters*, **70** (5), 532–534, URL <http://www.ifjf.uib.no/seism0/s0Rware.html>.
- Heidbach, O., and J. Reinecker, 2013: Arbeitsbericht NAB 12-05 - Analyse des rezenten Spannungsfelds der Nordschweiz. Tech. Rep. April, National Genossenschaft für die Lagerung radioaktiver Abfälle.
- Heidbach, O., and Coauthors, 2018: The World Stress Map database release 2016: Crustal stress pattern across scales. *Tectonophysics*, **744**, 484–498, doi:10.1016/j.tecto.2018.07.007.
- Henrich, R., and H. Zankl, 1981: Die Geologie des Hochstaufenmassives in den Nördlichen Kalkalpen. *Verhandlungen der Geologischen Bundesanstalt (Wien)*, **1981**, 31–57.
- Hergert, T., O. Heidbach, K. Reiter, S. B. Giger, and P. Marschall, 2015: Stress field sensitivity analysis in a sedimentary sequence of the Alpine foreland, northern Switzerland. *Solid Earth*, **6** (2), 533–552, doi:10.5194/se-6-533-2015.
- Houlié, N., J. Woessner, D. Giardini, and M. Rothacher, 2018: Lithosphere strain rate and stress field orientations near the Alpine arc in Switzerland. *Scientific Reports*, **8** (1), doi:10.1038/s41598-018-20253-z.

- Hubbert, M. K., and W. W. Rubey, 1959: Role of fluid pressure in mechanics of overthrust faulting: I. Mechanics of fluid-filled porous solids and its application to overthrust faulting. *GSA Bulletin*, **70** (2), 115–166, doi:10.1130/0016-7606(1959)70[115:rofpim]2.0.co;2.
- Huguenin, S., 2020: Désobstruction de la Source du Torrent. *Cavernes*, **1**, 14–26.
- Husen, S., C. Bachmann, and D. Giardini, 2007: Locally triggered seismicity in the central Swiss Alps following the large rainfall event of August 2005. *Geophysical Journal International*, **171** (3), 1126–1134, doi:10.1111/j.1365-246X.2007.03561.x.
- Husen, S., E. Kissling, N. Deichmann, S. Wiemer, D. Giardini, and M. Baer, 2003: Probabilistic earthquake location in complex three-dimensional velocity models: Application to Switzerland. *Journal of Geophysical Research*, **108** (B2), doi:10.1029/2002jb001778.
- Jaeger, J. C., N. G. Cook, and R. Zimmerman, 2009: *Fundamentals of rock mechanics*. Wiley.
- Jeannin, P. Y., 2001: Modeling flow in phreatic and epiphreatic karst conduits in the Hölloch cave (Muotatal, Switzerland). *Water Resources Research*, **37** (2), 191–200, doi:10.1029/2000WR900257.
- Jolivet, L., and C. Faccenna, 2000: Mediterranean extension and the Africa-Eurasia collision. *Tectonics*, **19** (6), 1095–1106, doi:10.1029/2000TC900018.
- Jouves, J., S. Viseur, B. Arfib, C. Baudement, H. Camus, P. Collon, and Y. Guglielmi, 2017: Speleogenesis, geometry, and topology of caves: A quantitative study of 3D karst conduits. *Geomorphology*, **298** (January), 86–106, doi:10.1016/j.geomorph.2017.09.019.
- Julian, M., and J. Nicod, 1984: Paléokarsts et paléo-géomorphologie néogènes des Alpes Occidentales et régions adjacentes. *Karstologia : revue de karstologie et de spéléologie physique*, **4** (1), 11–18, doi:10.3406/karst.1984.939.
- Kakurina, M., 2019: Mechanics of fault reactivation and its application to stress measurements. Ph.D. thesis, Université de Neuchâtel, Neuchâtel.
- Kangi, A., and N. Heidari, 2008: Reservoir-induced seismicity in Karun III dam (Southwestern Iran). *Journal of Seismology*, **12** (4), 519–527, doi:10.1007/s10950-008-9104-4.
- Keranen, K. M., H. M. Savage, G. A. Abers, and E. S. Cochran, 2013: Potentially induced earthquakes in Oklahoma, USA: Links between wastewater injection and the 2011 Mw 5.7 earthquake sequence. *Geology*, **41** (6), 699–702, doi:10.1130/G34045.1.
- Keranen, K. M., M. Weingarten, G. A. Abers, B. A. Bekins, and S. Ge, 2014: Sharp increase in central Oklahoma seismicity since 2008 induced by massive wastewater injection. *Science*, **345** (6195), 448–451, doi:10.1126/science.1255802.
- Kettermann, M., L. Smeraglia, C. K. Morley, C. Von Hagke, and D. C. Tanner, 2020: Fault sealing. *Understanding Faults: Detecting, Dating, and Modeling*, Elsevier, 283–350, doi:10.1016/B978-0-12-815985-9.00008-4.

- King, G. C., R. S. Stein, and Jian Lin, 1994: Static stress changes and the triggering of earthquakes. *Bulletin - Seismological Society of America*, **84** (3), 935–953, doi:10.1016/0148-9062(95)94484-2.
- Kinoshita, C., Y. Kano, and H. Ito, 2015: Shallow crustal permeability enhancement in central Japan due to the 2011 Tohoku earthquake. *Geophysical Research Letters*, **42** (3), 773–780, doi:10.1002/2014GL062792, URL <http://doi.wiley.com/10.1002/2014GL062792>.
- Kiraly, L., 1973: Notice explicative de la carte hydrogéologique du Canton de Neuchâtel. *Bulletin de la société Neuchâteloise des sciences naturelles*.
- Király, L., 1994: Groundwater flow in fractured rocks: models and reality. 14. *Mintrop Seminar über Interpretationsstrategien in Exploration und Produktion*, Ruhr Universität Bochum, 1–21.
- Király, L., 1998: Modelling karst aquifers by the combined discrete channel and continuum approach. *Bulletin du Centre d'Hydrogéologie de l'Université de Neuchâtel*, **15**, 77–98.
- Kovács, A., P. Perrochet, L. Király, and P.-Y. Jeannin, 2005: A quantitative method for the characterisation of karst aquifers based on spring hydrograph analysis. *Journal of Hydrology*, **303**, 152–164, doi:10.1016/j.jhydrol.2004.08.023.
- Kraft, T., J. Wassermann, and H. Igel, 2006a: High-precision relocation and focal mechanism of the 2002 rain-triggered earthquake swarms at Mt Hochstaufen, SE Germany. *Geophysical Journal International*, **167** (3), 1513–1528, doi:10.1111/j.1365-246X.2006.03171.x.
- Kraft, T., J. Wassermann, E. Schmedes, and H. Igel, 2006b: Meteorological triggering of earthquake swarms at Mt. Hochstaufen, SE-Germany. *Tectonophysics*, **424** (3-4), 245–258, doi:10.1016/j.tecto.2006.03.044.
- Kraft, T., and Coauthors, 2009: Enhanced geothermal systems: Mitigating risk in urban areas. *Eos*, **90** (32), 273–274, doi:10.1029/2009EO320001.
- Lanza, F., T. Diehl, N. Deichmann, T. Kraft, C. Nussbaum, S. Schefer, and S. Wiemer, 2022: The Saint-Ursanne earthquakes of 2000 revisited: evidence for active shallow thrust faulting in the Jura fold-and-thrust belt. *Swiss Journal of Geosciences*, **115** (1), doi:10.1186/s00015-021-00400-x.
- Leitner, C., and C. Spötl, 2017: *The Eastern Alps*. Elsevier Inc., 467–482 pp., doi:10.1016/b978-0-12-809417-4.00022-7, URL <http://dx.doi.org/10.1016/B978-0-12-809417-4.00022-7>.
- Lesparre, N., and Coauthors, 2017: New insights on fractures deformation from tiltmeter data measured inside the Fontaine de Vaucluse karst system. *Geophysical Journal International*, **208** (3), 1389–1402, doi:10.1093/gji/ggw446.
- Lienert, B. R., E. Berg, and L. N. Frazer, 1986: HYPOCENTER: An earthquake location method using centered, scaled, and adaptively damped least squares. *Bulletin of the Seismological Society of America*, **76** (3), 771–783.

- Lockner, D. A., and N. M. Beeler, 1999: Premonitory slip and tidal triggering of earthquakes. *Journal of Geophysical Research: Solid Earth*, **104** (B9), 20 133–20 151, doi:10.1029/1999jb900205.
- Lockner, D. A., and N. M. Beeler, 2002: Rock Failure and Earthquakes. *International handbook of earthquake and engineering seismology*, Vol. 81, IASPEI, 505–537, doi:10.1016/S0074-6142(02)80235-2.
- Lund, B., and R. Slunga, 1999: Stress tensor inversion using detailed microearthquake information and stability constraints: Application to Ölfus in southwest Iceland. *Journal of Geophysical Research: Solid Earth*, **104** (B7), 14 947–14 964, doi:10.1029/1999jb900111.
- Madritsch, H., F. Preusser, O. Fabbri, V. Bichet, F. Schlunegger, and S. M. Schmid, 2010: Late Quaternary folding in the Jura Mountains: Evidence from syn-erosional deformation of fluvial meanders. *Terra Nova*, **22** (2), 147–154, doi:10.1111/j.1365-3121.2010.00928.x.
- Manga, M., C. Y. Wang, and M. Shirzaei, 2016: Increased stream discharge after the 3 September 2016 Mw 5.8 Pawnee, Oklahoma earthquake. *Geophysical Research Letters*, **43** (22), 588–11, doi:10.1002/2016GL071268.
- Matonti, C., J. Lamarche, Y. Guglielmi, and L. Marié, 2012: Structural and petrophysical characterization of mixed conduit/seal fault zones in carbonates: Example from the Castellás fault (SE France). *Journal of Structural Geology*, **39**, 103–121, doi:10.1016/j.jsg.2012.03.003, URL <http://dx.doi.org/10.1016/j.jsg.2012.03.003>.
- Matthey, B., 1976: Hydrogéologie des bassins de la Serrière et du Seyon. Ph.D. thesis, Université de Neuchâtel, Neuchâtel.
- Métivier, L., O. de Viron, C. P. Conrad, S. Renault, M. Diamant, and G. Patau, 2009: Evidence of earthquake triggering by the solid earth tides. *Earth and Planetary Science Letters*, **278** (3–4), 370–375, doi:10.1016/j.epsl.2008.12.024.
- Michael, A. J., 1984: Determination of stress from slip data: faults and folds. *Journal of Geophysical Research: Solid Earth*, **89** (B13), 11 517–11 526.
- Michael, A. J., 1987: Use of focal mechanisms to determine stress: a control study. *Journal of Geophysical Research: Solid Earth*, **92** (B1), 357–368.
- Miller, S. A., 2008: Note on rain-triggered earthquakes and their dependence on karst geology. *Geophysical Journal International*, **173** (1), 334–338, doi:10.1111/j.1365-246X.2008.03735.x.
- Miller, S. A., C. Collettini, L. Chiaraluce, M. Cocco, M. Barchi, and B. J. P. Kaus, 2004: Aftershocks driven by a high-pressure CO₂ source at depth. Tech. rep. URL www.nature.com/nature.
- Morris, A., D. A. Ferrill, and D. B. Henderson, 1996: Slip-tendency analysis and fault reactivation. *Geology*, **24** (3), 275–278, doi:10.1130/0091-7613(1996)024<0275:STAAFR>2.3.CO;2.

- Mosar, J., 1999: Present-day and future tectonic underplating in the western Swiss Alps: reconciliation of basement/wrench-faulting and décollement folding of the Jura and Molasse basin in the Alpine foreland. Tech. rep., 143–155 pp. URL www.elsevier.com/locate/epsl.
- Muço, B., 1995: The seasonality of Albanian earthquakes and cross-correlation with rainfall. *Physics of the Earth and Planetary Interiors*, **88**, 285–291.
- Nagra, 1996: Technischer Bericht 96-01: Geosynthese Wellenberg 1996 - Ergebnisse der Untersuchungsphase I und II. *Technischer Bericht 96-01*, URL [https://www.nagra.ch/data/documents/database/dokumente/\\$default/DefaultFolder/Publikationen/NTBs1994-2000/d_ntb96-01_Textband.pdf](https://www.nagra.ch/data/documents/database/dokumente/$default/DefaultFolder/Publikationen/NTBs1994-2000/d_ntb96-01_Textband.pdf).
- Ndiaye, M., N. Clerc, G. Gorin, S. Girardclos, and J. Fiore, 2014: Lake Neuchâtel (Switzerland) seismic stratigraphic record points to the simultaneous Würmian deglaciation of the Rhône Glacier and Jura Ice Cap. *Quaternary Science Reviews*, **85** (February), 1–19, doi:10.1016/j.quascirev.2013.11.017.
- Ohnaka, M., Y. Kuwahara, K. Yamamoto, and T. Hirasawa, 2013: Dynamic Breakdown Processes and the Generating Mechanism for High-Frequency Elastic Radiation During Stick-Slip Instabilities. *Earthquake Source Mechanics*, S. Das, J. Boatwright, and C. H. Scholz, Eds., Vol. 37, American Geophysical Union, 13–24, doi:10.1029/gm037p0013.
- Ottensmöller, L., P. Voss, and J. Havskov, 2017: SEISAN Earthquake Analysis Software for windows, solaris, linux and macosx. (February).
- Ozawa, S., M. Murakami, M. Kaidzu, T. Tada, T. Sagiya, Y. Hatanaka, H. Yarai, and T. Nishimura, 2002: Detection and monitoring of ongoing aseismic slip in the Tokai region, central Japan. *Science*, **298** (5595), 1009–1012, doi:10.1126/science.1076780.
- Palmer, A. N., 1991: Origin and morphology of limestone caves. *Geological Society of America Bulletin*, **103** (1), 1–21, doi:10.1130/0016-7606(1991)103<0001:OAMOLC>2.3.CO;2.
- Parotidis, M., S. A. Shapiro, and E. Rothert, 2004: Back front of seismicity induced after termination of borehole fluid injection. *Geophysical Research Letters*, **31** (2), 1–5, doi:10.1029/2003GL018987.
- Pfiffner, O. A., 2011: *Structural Map of the Helvetic Zone of the Swiss Alps including Vorarlberg (Austria) and Haute Savoie (France), 1:100 000*. Geological Special Map 128, Swiss Federal Office of Topography swisstopo.
- Poulos, H., and E. Davis, 1974: *Elastic solutions for soil and rock mechanics*. John Wiley & Sons, Ltd.
- Preisig, G., and Coauthors, 2016: Development of connected permeability in massive crystalline rocks through hydraulic fracture propagation and shearing accompanying fluid injection. *Crustal Permeability*, 335–352, doi:10.1002/9781119166573.ch26.

- Rabin, M., C. Sue, A. Walpersdorf, P. Sakic, J. Albaric, and B. Fores, 2018: Present-Day Deformations of the Jura Arc Inferred by GPS Surveying and Earthquake Focal Mechanisms. *Tectonics*, **37** (10), 3782–3804, doi:10.1029/2018TC005047.
- Rakhmanov, E. A., E. B. Saff, and Y. M. Zhou, 1994: MINIMAL DISCRETE ENERGY ON THE SPHERE. *Mathematical Research Letters*, **1**, 647–662.
- Reiter, K., O. Heidbach, B. Müller, J. Reinecker, and T. Röckel, 2016: Stress Map Germany 2016. GFZ Data Services, 1 pp., doi:10.5880/WSM.Germany2016.
- Rey, T., and J. Mosar, 2015: Etude Structurale du Jura Neuchâtelois. *Université de Fribourg*.
- Rice, J. R., and M. P. Cleary, 1976: Some basic stress diffusion solutions for fluid-saturated elastic porous media with compressible constituents. *Reviews of Geophysics*, **14** (2), 227–241, doi:10.1029/RG014i002p00227.
- Rigo, A., N. Béthoux, F. Masson, and J. F. Ritz, 2008: Seismicity rate and wave-velocity variations as consequences of rainfall: The case of the catastrophic storm of September 2002 in the Nîmes Fault region (Gard, France). *Geophysical Journal International*, **173** (2), 473–482, doi:10.1111/j.1365-246X.2008.03718.x.
- Rime, V., A. Sommaruga, M. Schori, and J. Mosar, 2019: Tectonics of the Neuchâtel Jura Mountains: insights from mapping and forward modelling. *Swiss Journal of Geosciences*, **112** (2-3), 563–578, doi:10.1007/s00015-019-00349-y.
- Roeloffs, E. A., 1988: Fault stability changes induced beneath a reservoir with cyclic variations in water level. *Journal of Geophysical Research*, **93** (B3), 2107, doi:10.1029/jb093ib03p02107.
- Rojstaczer, S., S. Wolf, and R. Michel, 1995: Permeability enhancement in the shallow crust as a cause of earthquake-induced hydrological changes. *Nature*, **373**, 237–239.
- Roth, P., N. Pavoni, and N. Deichmann, 1992: Seismotectonics of the eastern Swiss Alps and evidence for precipitation-induced variations of seismic activity. *Tectonophysics*, **207** (1-2), 183–197, doi:10.1016/0040-1951(92)90477-N.
- Saar, M. O., and M. Manga, 2003: Seismicity induced by seasonal groundwater recharge at Mt. Hood, Oregon. *Earth and Planetary Science Letters*, **214** (3-4), 605–618, doi:10.1016/S0012-821X(03)00418-7.
- Savage, W. Z., H. S. Swolfs, and B. Amadei, 1992: On the state of Stress in the Near-surface of the Earth's Crust. *PAGEOPH*, **138** (2).
- Schlupp, A., G. Clauzon, and J. P. Avouac, 2001: Mouvement post-messinien sur la faille de Nîmes: Implications pour la sismotectonique de la Provence. *Bulletin de la Societe Geologique de France*, **172** (6), 697–711, doi:10.2113/172.6.697.
- Schmid, S. M., B. Fügenschuh, E. Kissling, and R. Schuster, 2004: Tectonic map and overall architecture of the Alpine orogen. *Eclogae Geologicae Helvetiae*, **97** (1), 93–117, doi:10.1007/s00015-004-1113-x.

- Schmitt, D. R., C. A. Currie, and L. Zhang, 2012: Crustal stress determination from boreholes and rock cores: Fundamental principles. *Tectonophysics*, **580 (Supplement C)**, 1–26, doi:10.1016/j.tecto.2012.08.029, URL <http://www.sciencedirect.com/science/article/pii/S0040195112005185>.
- Schmitt, D. R., and B. Haimson, 2017: Hydraulic fracturing stress measurements in deep holes. *Rock Mechanics and Engineering Volume 1*, doi:10.1201/9781315364261-7, URL <https://www.taylorfrancis.com/>.
- Scholz, C. H., 1998: Earthquakes and friction laws. *NATURE*, **391**.
- Scholz, C. H., Y. J. Tan, and F. Albino, 2019: The mechanism of tidal triggering of earthquakes at mid-ocean ridges. *Nature Communications*, **10 (1)**, 1–7, doi:10.1038/s41467-019-10605-2.
- Schori, M., J. Mosar, and G. Schreurs, 2015: Multiple detachments during thin-skinned deformation of the Swiss Central Jura: a kinematic model across the Chasseral. *Swiss Journal of Geosciences*, **108 (2-3)**, 327–343, doi:10.1007/s00015-015-0196-x.
- Schwarzmann, A., 1996: Untersuchung der seismischen Aktivität im Raum Bad Reichenhall. Ph.D. thesis, University of Munich.
- Schweigl, J., and F. Neubauer, 1997: Structural evolution of the central Northern Calcareous Alps: Significance for the Jurassic to Tertiary geodynamics in the Alps. *Eclogae Geologicae Helvetiae*, **90 (2)**, 303–323.
- Seront, B., T. F. Wong, J. S. Caine, C. B. Forster, R. L. Bruhn, and J. T. Fredrich, 1998: Laboratory characterization of hydromechanical properties of a seismogenic normal fault system. *Journal of Structural Geology*, **20 (7)**, 865–881, doi:10.1016/S0191-8141(98)00023-6.
- Shapiro, S. A., and C. Dinske, 2021: Stress Drop, Seismogenic Index and Fault Cohesion of Fluid-Induced Earthquakes. *Rock Mechanics and Rock Engineering*, **54 (10)**, 5483–5492, doi:10.1007/s00603-021-02420-3.
- Shi, Z., G. Wang, M. Manga, and C. Y. Wang, 2015: Continental-scale water-level response to a large earthquake. *Geofluids*, **15 (1-2)**, 310–320, doi:10.1111/gfl.12099.
- Sibson, R., K. Aki, M. Ohnaka, L. Ruff, and P. Spudich, 1992: Implications of fault-valve behaviour for rupture nucleation and recurrence. Tech. rep., 283–293 pp.
- Sibson, R. H., 1985: A note on fault reactivation. *Journal of Structural Geology*, **7 (6)**, 751–754, doi:10.1016/0191-8141(85)90150-6.
- Sjöberg, J., R. Christiansson, and J. A. Hudson, 2003: ISRM suggested methods for rock stress estimation-part 2: Overcoring methods. Elsevier BV, 999–1010 pp., doi:10.1016/j.ijrmms.2003.07.012.
- Sommaruga, A., 1997: *Geology of the Central Jura and the Molasse Basin*, Vol. 12. Société neuchâteloise des Sciences naturelles, Neuchâtel, 195 pp.

- Sommaruga, A., 2011: From the central Jura Mountains to the Molasse Basin (France and Switzerland). *Bulletin fuer Angewandte Geologie*, **16** (2), 63–75.
- Stephansson, O., and A. Zang, 2012: ISRM suggested methods for rock stress estimation-part 5: Establishing a model for the in situ stress at a given site. *Rock Mechanics and Rock Engineering*, **45** (6), 955–969, doi:10.1007/s00603-012-0270-x.
- Stokes, S. M., S. Ge, M. R. Brown, E. A. Menezes, A. F. Sheehan, and K. F. Tiampo, 2023: Pore Pressure Diffusion and Onset of Induced Seismicity. *Journal of Geophysical Research: Solid Earth*, **128** (3), doi:10.1029/2022JB026012.
- Swiatlowski, J. L., 2020: Understanding Fault Creep From the Macroscopic to the Microscopic Scale. Ph.D. thesis, UC Riverside.
- Talwani, P., 1997: On the Nature of Reservoir-induced Seismicity. *Pure and applied geophysics*, **150** (3), 473–492, doi:10.1007/s000240050089.
- Talwani, P., and S. Acree, 1984: Pore pressure diffusion and the mechanism of reservoir-induced seismicity. *Pure and Applied Geophysics PAGEOPH*, **122** (6), 947–965, doi:10.1007/BF00876395, URL <https://link.springer.com/article/10.1007/BF00876395>.
- Terrier, M., G. Courrioux, G. Martelet, A. Bitri, and F. Hanot, 2007: Etude de la faille de Salon-Cavaillon (région PACA, France) - Rapport final. Tech. rep., BRGM RP-55989-Fr, 177 p., 66 fig., 3 ann. pp.
- Tollmann, A., 1976: *Der Bau der Nördlichen Kalkalpen. Orogene Stellung und regionale Tektonik*, Vol. 3. Deuticke, Wien.
- Tolstoy, M., F. L. Vernon John A Orcutt Frank K Wyatt, and I. M. Green, 2002: Breathing of the seafloor: Tidal correlations of seismicity at Axial volcano. Tech. Rep. 6, 503 pp.
- Townend, J., and M. D. Zoback, 2000: How faulting keeps the crust strong. *Geology*, **28** (5), 399–402, doi:10.1130/0091-7613(2000)028<0399:HFKTCS>2.3.CO;2.
- Trnkoczy, A., 2009: Understanding and parameter setting of STA/LTA trigger algorithm. *New Manual of Seismological Observatory Practice (NMSOP-2)*, IASPEI.
- Turner, F., 1962: "Compression" and "tension" axes determined from {0112} twinning in calcite. *Journal of Geophysical Research*, **67**.
- Ustaszewski, K., and S. M. Schmid, 2007: Neotectonic activity in the Upper Rhine Graben - Jura mountains junction (North-Western Switzerland and adjacent France). *Bulletin fuer Angewandte Geologie*, **12** (1), 3–19.
- Valley, B., and K. F. Evans, 2019: Stress magnitudes in the Basel enhanced geothermal system. *International Journal of Rock Mechanics and Mining Sciences*, **118**, 1–20, doi:10.1016/j.ijrmms.2019.03.008.
- Vavryčuk, V., 2014: Iterative joint inversion for stress and fault orientations from focal mechanisms. *Geophysical Journal International*, **199** (1), 69–77, doi:10.1093/gji/ggu224.

- Vouillamoz, N., G. H. Wust Bloch, M. Abednego, and J. Mosar, 2016: Optimizing event detection and location in low-seismicity zones: Case study from Western Switzerland. *Bulletin of the Seismological Society of America*, **106** (5), 2023–2036, doi:10.1785/0120160029.
- Waldhauser, F., and W. L. Ellsworth, 2000: A Double-Difference Earthquake Location Algorithm: Method and Application to the Northern Hayward Fault, California. *Bulletin of the Seismological Society of America*, **90**, 1353–1368.
- Wallace, R. E., 1951: Geometry of shearing stress and relation to faulting. *The Journal of Geology*, **59** (2), 118–130, URL <http://www.journals.uchicago.edu/t-and-c>.
- Wang, C. Y., L. P. Wang, M. Manga, C. H. Wang, and C. H. Chen, 2013: Basin-scale transport of heat and fluid induced by earthquakes. *Geophysical Research Letters*, **40** (15), 3893–3897, doi:10.1002/grl.50738.
- Weiss, J., V. Pellissier, D. Marsan, L. Arnaud, and F. Renard, 2016: Cohesion versus friction in controlling the long-term strength of a self-healing experimental fault. *Journal of Geophysical Research: Solid Earth*, **121** (12), 8523–8547, doi:10.1002/2016JB013110.
- Weiss, L., 1954: A Study of tectonic style: Structural investigation of a marble quartzite complex in southern California. *University of California publications in geological sciences*, **30**, 1–102.
- Wiemer, S., D. Giardini, D. Fäh, N. Deichmann, and S. Sellami, 2009: Probabilistic seismic hazard assessment of Switzerland: Best estimates and uncertainties. *Journal of Seismology*, **13** (4), 449–478, doi:10.1007/s10950-008-9138-7.
- Wildberger, A., M. A. Geyh, U. Groner, P. Häuselmann, F. Heller, and M. Ploetze, 2010: Dating speleothems from the Silberen Cave System and surrounding areas: Speleogenesis in the Muota Valley (Central Switzerland). *Zeitschrift für Geomorphologie*, **54** (SUPPL. 2), 307–328, doi:10.1127/0372-8854/2010/0054S2-0016.
- Wolf, L. W., C. A. Rowe, and R. B. Horner, 1997: Periodic seismicity near Mt. Ogden on the Alaska-British Columbia border: A case for hydrologically triggered earthquakes? *Bulletin of the Seismological Society of America*, **87** (6), 1473–1483.
- Yamaji, A., 2003: Are the solutions of stress inversion correct? Visualization of their reliability and the separation of stresses from heterogeneous fault-slip data. *Journal of Structural Geology*, **25**, 241–252, URL www.elsevier.com/locate/jsg.
- Zaccagnino, D., L. Telesca, and C. Doglioni, 2022: Correlation between seismic activity and tidal stress perturbations highlights growing instability within the brittle crust. *Scientific Reports*, **12** (1), doi:10.1038/s41598-022-11328-z.
- Zang, A., and O. Stephansson, 2010: *Stress Field of the Earth's Crust*. 9, Springer Netherlands, Dordrecht, doi:10.1007/978-1-4020-8444-7, URL http://www.elsevier.com/locate/scphhttp://link.springer.com/10.1007/978-1-4020-8444-7_9.

- Zoback, M. D., and S. M. Gorelick, 2012: Earthquake triggering and large-scale geologic storage of carbon dioxide. *Proceedings of the National Academy of Sciences*, **109** (26), 10 164–10 168, doi:10.1073/pnas.1202473109.
- Zoback, M. D., and Coauthors, 2003: Determination of stress orientation and magnitude in deep wells. *International Journal of Rock Mechanics and Mining Sciences*, **40** (7-8), 1049–1076, doi:10.1016/j.ijrmms.2003.07.001.
- Zoback, M. L., and Coauthors, 1989: Global patterns of tectonic stress. *Nature*, **341** (6240), 291–298, doi:10.1038/341291a0, URL <https://doi.org/10.1038/341291a0>.



This study assesses the criticality of faults in karstic regions by analyzing hydrogeological and microseismic data, aiming to enhance the understanding of seismic activity and groundwater interactions. These objectives are prompted by concerns regarding induced seismicity and its associated risks in projects involving, for example, deep geothermal energy exploitation, CO₂ sequestration, or radioactive waste disposal.

In this study, seismic and hydrogeological data are integrated to identify whether seismicity occurs after significant recharge periods in the Neuchâtel area. Considering different mechanisms (e.g., direct hydrogeological connection, pore pressure diffusion, poro-elastic deformation), 1D analytical solutions are used to determine the pore pressure changes following the recharge events and eventually allow identifying the most impacting mechanism. The results of this study emphasize the importance of considering the hydrogeological context, especially in a karstic environment where the channeling effect of karst conduits has a major influence on pore pressure changes at depth.

